VOL. 249 JANUARY 1951 No. 1

American Journal of Science

Established in 1818 by Benjamin Silliman

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Published monthly. Eight dollars a year. Seventy-five cents a copy.
Foreign, eight dollars sixty cents. Canada, eight dollars thirty-five cents.

AMERICAN JOURNAL OF SCIENCE

Established in 1818 by Benjamin Silliman

THE FIRST SCIENTIFIC JOURNAL IN THE UNITED STATES

DEVOTED TO THE GEOLOGICAL SCIENCES AND TO RELATED FIELDS

Editor: CHESTER RAY LONGWELL
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Entered as second-class matter at the Post Office at New Haven, Conn., under the Act of March 3, 1879. Published monthly at Sterling Tower, Yale University, New Haven, Conn.

Subscription rate $8 per year (75 cents a number). Student rate $5 per year (application blanks available on request). Postage prepaid to the United States and Central and South America; 35 cents per year to Canada; 60 cents per year elsewhere.

The JOURNAL completed its *first* series of 50 volumes in 1845, its *second* series of 50 volumes in 1870, its *third* series of 50 volumes in 1895, its *fourth* series of 50 volumes in 1920, its *fifth* series of 36 volumes in 1938. Since 1938 the numbering by series has been discontinued, and volumes are annual and bear the whole numbers as of 1818, that for 1939 being volume 237.

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REFERENCE

Lane, B. H., 1935. Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C., 25¢.)

[American Journal of Science, Vol. 249, January, 1951, Pp. 1-30]

American Journal of Science

JANUARY 1951

STRUCTURE OF THE HOT SPRINGS WINDOW, MADISON COUNTY, NORTH CAROLINA*

STEVEN S. ORIEL

ABSTRACT. The Hot Springs window, in western Madison County, North Carolina, and eastern Cocke County, Tennessee, lies near the northwest margin of the Blue Ridge physiographic province.

The rocks within the window include a pre-Cambrian crystalline complex and a thick overlying sequence of sedimentary rocks, which is composed, in ascending order, of about 5200 feet of clastic rocks of unknown age (Snowbird and Sandsuck formations), at least 4300 feet of clastic rocks of Early Cambrian age (Unicoi, Hampton, and Erwin formations), and an unknown thickness of carbonate and fine-grained clastic rocks of Early and Middle Cambrian age (Shady, Rome, and Honaker formations). The rocks framing the window belong to the same formations, up to and including the Unicoi formation.

The Hot Springs window is unusual in that it is framed by not one, but four different thrust sheets. The rocks at and near the thrust faults are considerably fractured and mylonitized in most exposures. Where granitoid rocks of two thrust sheets are adjacent to each other, mylonite belts are useful in tracing the intervening fault. Mylonitized pre-Cambrian crystalline rocks, however, are difficult to distinguish from mylonitized Snowbird arkose in many places. The fractured and mylonitized nature of the rocks along the contact between the Snowbird formation and the pre-Cambrian crystalline complex within the window suggest that this contact also may be a thrust surface, rather than an unconformity. If this contact is a fault, then the pre-Cambrian crystalline rocks adjacent to it occur in a small window within the Hot Springs window.

The Hot Springs window may be compared with a few other windows with multiple faulting. Both the Hot Springs window and the nearby Mountain City window are analogous in certain respects to the Engadine window of the Central Alps. All are framed by more than one thrust fault and are therefore eyelid windows. The sequence of tectonic events which led to the formation of the Hot Springs structure may have been comparable to that of the Engadine window, as interpreted by Sander.

* Part of a dissertation submitted to the faculty of Yale University in partial fulfilment of the requirements for the degree of Doctor of Philosophy. Published by permission of the State Geologist, North Carolina Department of Conservation and Development.

INTRODUCTION

Location.—The Hot Springs area lies in the western part of Madison County, North Carolina, and in the eastern part of Cocke County, Tennessee (fig. 1). The area covered by the geologic mapping on which this paper is based is elongate in an east-west direction with a length of about 10½ miles and a width of 5 miles, and thus includes about 55 square miles.

The Hot Springs area lies near the northwest margin of the Blue Ridge physiographic province. The mountains of the area are part of the Bald Mountains, which are in the Unaka Range of the Blue Ridge province. To the southeast lies the Blue Ridge Plateau on which Asheville, North Carolina is situated; to the northwest, the Appalachian Valley of Tennessee.

The Hot Springs window includes one of the few southeastern exposures of Paleozoic carbonate rocks in a province underlain predominantly by clastic, metamorphic, and granitic rocks. The presence of these carbonate rocks, as well as of North Carolina's only hot springs, and the occurrence of relatively good exposures have long attracted the interest of many geologists. However, previous work in the area was essentially of a reconnaissance nature whereas the present paper is based on a detailed study.

Previous work.—Keith (1904, p. 8) was the first to recognize the unusual nature of the structure around Hot Springs, N. C.; he mapped a single fault almost completely around the area of the Hot Springs window, but did not join the ends on the southeast side. Yet he believed that on the north side the thrust sheet had been overthrust from the north, and on the south side, from the south. "The area inclosed by the fault plane thus represents a downthrown mass upon which the adjoining rocks were piled high from all sides." The trace of the fault was described as an almost complete oval and "its planes, if extended upward, would almost unite in a dome".

The lithologic similarity between mylonitized granitoid rocks and sheared arkose (see below) led Keith to map the former as long, thin isoclinal synclines of Snowbird rocks in the Max Patch granite. As a result, several thrust faults were overlooked and the many "whiskers", or thin outcrop areas, of Snowbird rocks on Keith's map make the geological structure appear overly complex.

The Hot Springs area was first shown as a window on the

Geological Map of the United States (1932) but it was not described as such until 1944 (Stose and Stose, 1944, pp. 385-386). The window is described more fully, however, in a paper by Stose and Stose (1947) on the origin of the hot springs. In that paper they describe (pp. 624-625) the structural unit as " . . . the Hot Springs window of the Great Smoky overthrust block which has been domed and later eroded, exposing overridden rocks in an oval window". The long narrow isoclinal synclines of Snowbird rocks on Keith's map are interpreted by them as mylonite zones along subsidiary thrust faults which offset the trace of the Great Smoky thrust fault along the southern margin of the window. Their interpretation of the structure of the western part of the window is based upon Keith's map.

Stose and Stose, as well as Keith, were misled in their interpretation of the structure of the south-central part of the

Figure 1. Index map of western North Carolina and eastern Tennessee.

window by their belief that a belt of granitic rocks extends continuously from Bluff to the segment of the river between Stackhouse and Sandy Bottom; the present writer found that the western flank of Spring Creek Mountain, east of Bluff, is underlain by sedimentary rocks of the Snowbird formation (fig. 2).

In a manuscript soon to be published as a bulletin of the Tennessee Division of Geology, Ferguson (Ferguson and Jewell, in press) describes in detail the geology of the Del Rio district, Cocke County, Tennessee, including the westernmost portion of the Hot Springs area (fig. 1). Ferguson points out that the Hot Springs window is framed on the west by at least two thrust faults, and that one intersects and has overridden the other. His competent field mapping and detailed discussions of the structural and stratigraphic relations of the thrust blocks lying to the northwest of the Hot Springs area will be invaluable to future regional studies.

Field work.—The field study of the Hot Springs window by the author was made over a period of five months, between June and November, in 1948. Detailed stratigraphic studies were restricted to the rocks within the window. Readily traceable lithologic units, especially resistant quartzite beds, were utilized as formational boundaries and followed along their strikes. Discordant contacts between rocks of different lithologies were also mapped in detail.

The topographic base used throughout the study was that of the Tennessee Valley Authority and the U. S. Geological Survey prepared on a scale of 1:24,000 from aerial photographs.[1]

Acknowledgments.—The writer is grateful for the assistance of many individuals and particularly to John Rodgers for enthusiastic aid, advice, and helpful criticisms throughout the study. Field work was financed by the North Carolina Department of Conservation and Development from funds pledged for cooperative work with the Tennessee Valley Authority; J. L. Stuckey, state geologist, C. E. Hunter, Benjamin Gildersleeve, and the late H. S. Rankin of the T.V.A. were most cooperative. Office and laboratory work was financed

[1] Specific localities mentioned in this paper but not shown in figure 2 may be found on the following topographic sheets of the TVA-USGS series: Paint Rock (182NW), Lemon Gap 182SW), Hot Springs (182NE), Spring Creek (182SE), White Rock (191NW), and Marshall (191SW).

by the Shell Oil Company, Inc., in the form of a fellowship at Yale University during the academic year 1948-1949. J. B. Cathey served as a conscientious field assistant. H. W. Ferguson, assistant state geologist of Tennessee, supplied invaluable unpublished data on the Del Rio district, and made available his field maps of the Tennessee portion of the Hot Springs window. John Rodgers, C. R. Longwell, E. B. Knopf, P. B. King, J. B. Hadley, and E. L. Reed read the manuscript or portions thereof. The illustrations were drafted by Mrs. Marian Wheeler, of the Office of Geologic Cartography, U. S. Geological Survey. To all of these, and others, the writer wishes to express his gratitude.

STRATIGRAPHY

Inasmuch as the structural features of the Hot Springs window are the primary concern of the present paper, only a brief outline of the stratigraphy is presented here. The reader is referred to Bulletin 60 of the North Carolina Department of Conservation and Development (Oriel, 1950) for more detailed discussions of the stratigraphic relations on which the conclusions presented here are based.[2]

All the formations shown in table 1, from the pre-Cambrian crystalline complex up to and including the Honaker limestone, are represented within the Hot Springs window. The outcrop belts of the formations trend approximately east-west, with the youngest formations to the north (fig. 2). In the northern part of the window, the beds dip steeply to the north and are slightly overturned in places. Dips decrease southward within the window. Outside the window, only the Unicoi and older formations are present. Here too, the youngest formation, the Unicoi, is on the northwest and is succeeded by older formations to the southeast.

Although the Hot Springs area forms a part of that described in the Asheville folio (Keith, 1904), the names applied to the rocks in the present paper differ from most of those in Keith's report. The reasons for the changes in terminology are discussed in the aforementioned bulletin (Oriel, 1950).

In general, all but the lower three units used here correspond to those recently described in detail by geologists in eastern

[2] The bulletin also contains petrographic data and descriptions of the geography, topography, mineral resources, and hot springs of the area herein described.

TABLE 1.

Generalized section of the geologic formations in the Hot Springs area.

Age	Group	Formation	Member	Character	Thickness in ft.
Middle Cambrian		Honaker limestone		White, gray, and blue-gray limestone with silty and shaly laminae and concentrically banded chert nodules.	?
		Rome formation		Red, maroon, and brown siltstone and shale with interbedded light gray and blue-gray dolomite and greenish shale.	?
		Shady dolomite		Blue-gray, light gray, white and ribboned dolomite with some interbedded limestone.	1975 (max.)
Early Cambrian	Clastic Group of Lower Cambrian Series	Erwin formation		Dark green siltstone, silty and sandy shale, white vitreous quartzite, and purple ferruginous quartzite, with calcareous siltstone (Helenmode member) at the top.	1700-2100
		Hampton formation	Upper shale	Alternating dark green siltstone, silty and sandy shale, and thin-bedded quartzite.	200-300
			Middle quartzite	Crossbedded medium- to coarse-grained feldspathic quartzite and arkose with pebbly beds and vitreous quartzite.	500-750
			Lower shale	Interbedded dark green siltstone, argillaceous shale, and sandy and silty shale.	210-380
		Unicoi formation		Coarse feldspathic and vitreous quartzite, conglomerate (some with slate inclusions), siltstone, slate, and micaceous sandstone.	1400-2600
	Unconformity (?)				
Unknown	Clastic Group of Unknown Age	Sandsuck formation		Dark green and blue-green slate, shale, and well-laminated siltstone with gray sandy and black slaty bands; includes conglomerate and limestone lenses.	(about) 700
		Snowbird formation		Interbedded pebbly, vitreous, and feldspathic quartzites, arkose, shale, siltstone, slate, micaceous and calcareous sandstones, and mylonite.	1500-4600 (min.)
	Unconformity (?)				
Pre-Cambrian		Crystalline complex		Sericitized and epidotized granitoid, aplitic, pegmatitic, and basic rocks (undifferentiated in this report) and mylonite.	

Tennessee (King *et al.*, 1944; Rodgers, 1948; Ordway, unpublished; Ferguson and Jewell, in press). King (1949), however, places the writer's "clastic group of unknown age", *i.e.* the Snowbird and Sandsuck formations, in the late pre-Cambrian Ocoee series, whereas Stose and Stose (1949) retain the Sandsuck in the Lower Cambrian Chilhowee group and abandon the name Snowbird.[3]

The Hot Springs window contains a rather isolated inlier of relatively young carbonate and finely clastic rocks in a region where clastic and highly metamorphosed rocks predominate. Except for a few occurrences near the tops of thrust sheets elsewhere in the Unaka Range (Ferguson and Jewell, in press), outcrops of the Shady, Rome, and Honaker formations are restricted largely to the Valley and Ridge province.

The Hot Springs area also includes the southeasternmost exposures of known, relatively unmetamorphosed, Cambrian clastic rocks. These rocks lie along the projected axis of the Mountain City window (King *et al.*, 1944, pp. 10-13) which, except for the Grandfather Mountain window, also contains the southeasternmost-known exposures of relatively unmetamorphosed Cambrian sedimentary rocks.

Unlike greatly deformed areas in the Valley and Ridge province, the Hot Springs area includes exposures of true basement rocks (pre-Cambrian crystalline complex) which were involved in the thrusting.

STRUCTURE

General Features of Window

The Hot Springs window is 10.4 miles long, along its east-west axis, and has a maximum width of 6.4 miles in a north-south direction. However, the area within the window is somewhat less than 50 square miles.

The window is framed by not one but four different thrust sheets (fig. 3). Moreover, if the contact between the sedimentary and the crystalline rocks inside the window is a thrust fault, as the writer believes, then there are two major structural units within the window.

Other structural features in the area are closely related to the thrusting. The orientation of axial planes of folds near the faults suggest a genetic relation to the thrusting. Where

[3] This problem too is discussed more fully elsewhere (Oriel, 1950).

evident, shear planes in mylonite closely parallel trends of the thrust surfaces. Cleavage developed in the area may be attributable to the forces responsible for the thrust faults. Moreover, the mineralization in the area is closely related to the thrust faults—at least geographically if not genetically.

The structural features here described cannot be dated accurately from evidence in the Hot Springs area. Inasmuch as all the rocks have been affected, deformation is post-Honaker, or post-Middle Cambrian in age. Thrusting in the Southern Appalachians, northwest of the Piedmont, is generally assigned to the Late Paleozoic (Appalachian) orogeny, but the possibility that at least some may have taken place during earlier orogenic pulses is by no means eliminated.

Thrust Faults

Faults framing window.—The four thrust faults and thrust sheets that frame the Hot Springs window are shown in figure 2 and shown and named in figures 3 and 4.

The Mine Ridge thrust fault is the most evident and best-documented thrust fault in the Hot Springs area. The fault trace extends three-fourths of the distance around the window, beginning at Wolf Creek on the west and passing eastward and southward to Coley Gap on the south side of the window. The fault separates rocks of the Snowbird, Sandsuck, and Unicoi formations outside the window from the predominantly younger rocks within, and truncates younger formations inside the window. Only on the southeastern side of the window, where Snowbird rocks above the fault overlie pre-Cambrian crystalline rocks, do younger rocks overlie older. On the north, the Mine Ridge thrust fault dips gently northward, whereas on the south and east sides it dips steeply to the southeast.

The Mine Ridge thrust fault forms the base of the Del Rio thrust sheet, named by Ferguson (Ferguson and Jewell, in press) for the town of Del Rio, Cocke County, Tennessee. In the Hot Springs area, only the Unicoi, Sandsuck, and Snowbird formations are present in the Del Rio sheet. In the area mapped by Ferguson, however, other formations up to and including the Rome are present.

The Brushy Mountain thrust fault was named by Ferguson for exposures on Brushy Mountain, Cocke County, Tennessee. In the course of his investigation, Ferguson traced the fault

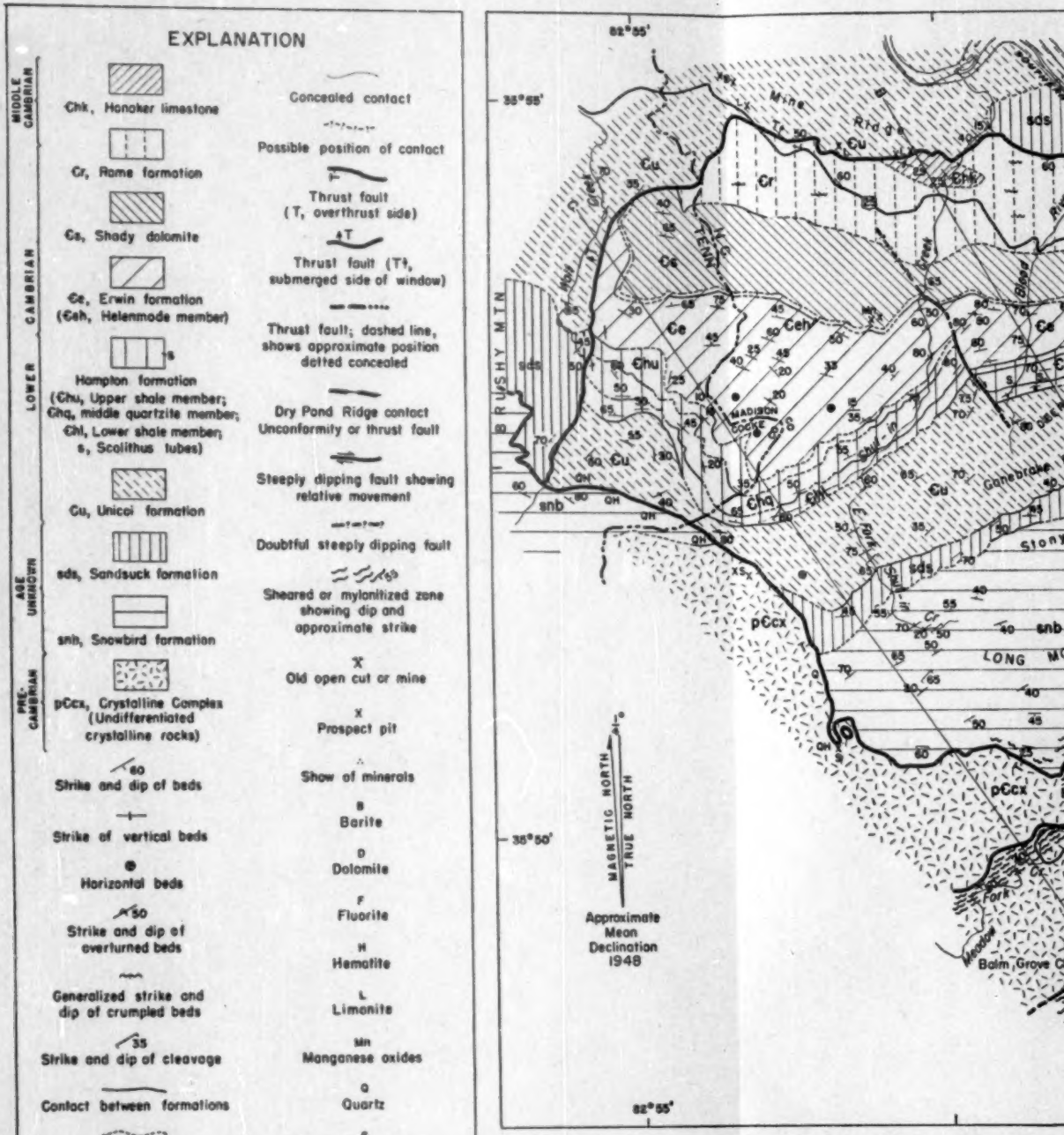
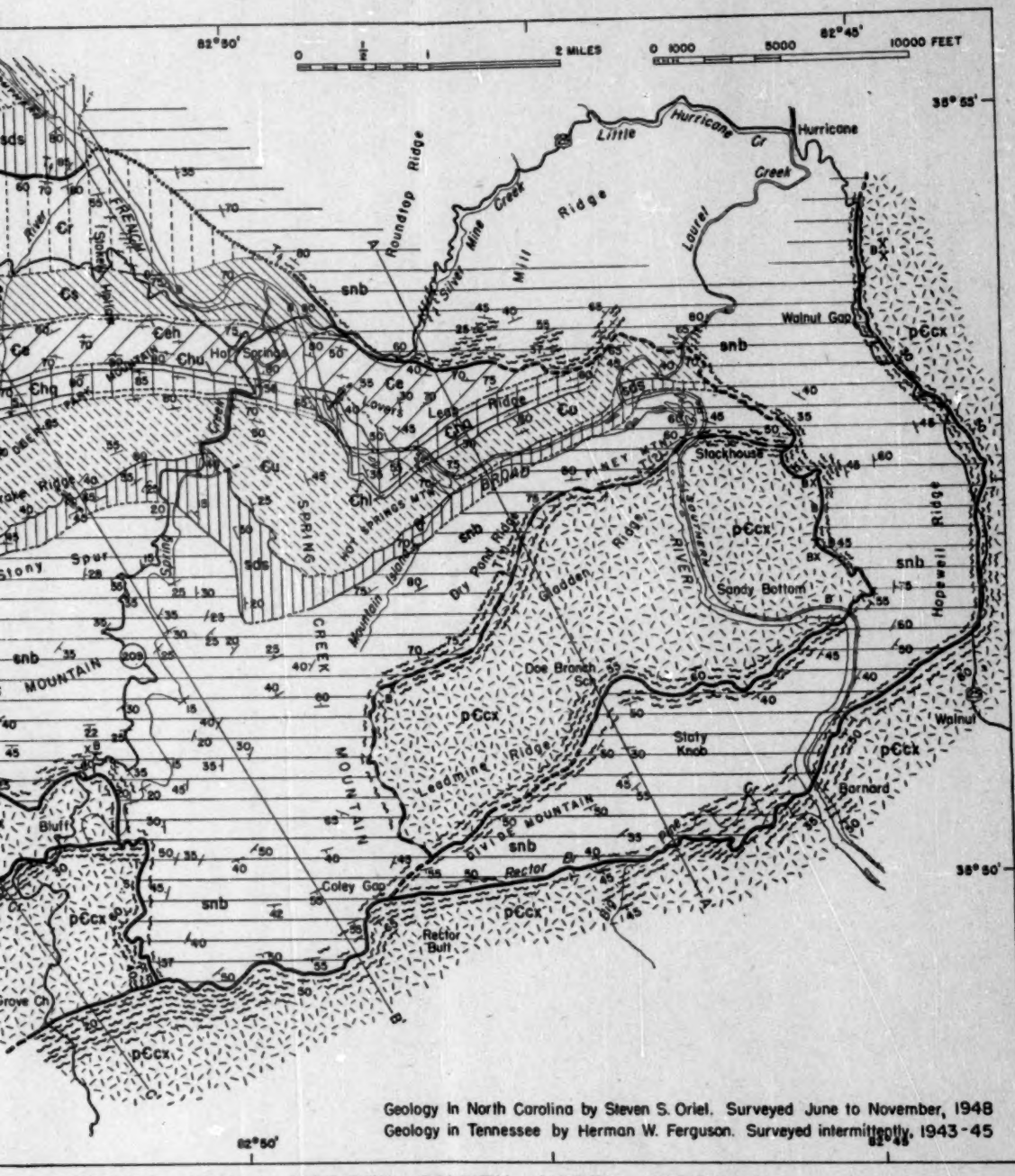


Figure 2. GEOLOGIC MAP OF HOT



OT SPRINGS AREA, MADISON COUNTY, NORTH CAROLINA

as far east as the old barite mine north of Bluff. The fault can be traced only a short distance beyond the mine to the point where it is truncated by the Meadow Fork thrust fault. As discovered by Ferguson, the trace of the Brushy Mountain thrust fault truncates the trace of the Mine Ridge fault at the west end of the window and it also truncates sedimentary contacts within the window. Almost everywhere along the fault in the area, sedimentary rocks within the window dip toward (under) the older crystalline rocks outside the window. In most places the fault dips to the south and southeast.

Above the Brushy Mountain fault is the Brushy Mountain thrust sheet. The sheet is present only in the southwestern part of the area mapped by the present writer and includes here only pre-Cambrian crystalline rocks and part of the Snowbird formation. The Brushy Mountain thrust sheet overlies the Del Rio thrust sheet as well as rocks within the window.

Only a small portion of the Meadow Fork thrust fault is exposed in the southern part of the Hot Springs area. The fault truncates the southeastern portion of the Brushy Mountain fault. It is probable that the fault is at least as extensive as the "synclinal" belt of "Snowbird" rocks shown on Keith's map (1904) extending six miles southwestward from Puncheon Camp Branch. Evidence is given below that this belt is a mylonite belt.

The Meadow Fork thrust sheet is here defined as the thrust sheet immediately overlying the Meadow Fork fault. The sheet consists entirely of pre-Cambrian crystalline rocks which are in part mylonitized. These rocks overlie rocks of the Brushy Mountain thrust sheet as well as those of the Snowbird formation within the window.

The Rector Branch thrust fault is named for the branch which follows the fault trace for over a mile in the south central part of the Hot Springs area. The trace of this thrust fault truncates the traces of both the Meadow Fork and Mine Ridge thrust faults, as well as bedding trends in the Snowbird rocks within the window. Projected dips, as well as several outcrops along U. S. Highway 25-70, show that the pre-Cambrian crystalline rocks above the fault overlie Snowbird rocks below the fault throughout much of its extent. The Rector Branch fault probably continues northward outside the area mapped in this study, at least as far as Little Hurricane Creek,

as shown on Keith's map (1904). On the southwest, it is possible that the fault continues across and beyond Hogback Mountain, but more detailed field work is necessary to test this view.

The hanging wall of the Rector Branch thrust fault is here designated the Rector Branch thrust sheet. Inasmuch as the Rector Branch fault was used as the approximate southern limit of the area covered by the writer, little can be said about this thrust sheet except that, where seen, it is made up of pre-Cambrian crystalline rocks. The thrust sheet directly overlies the Meadow Fork and Del Rio thrust sheets, as well as Snowbird rocks in the southern part of the window. Thus, it is the uppermost structural unit in the entire area.

Mylonite.—Immediately adjacent to the thrust faults throughout much of the area are thick zones of mylonite.[4] These proved to be useful in mapping the thrust faults, especially where granitoid rocks of the pre-Cambrian crystalline complex are in fault contact with other granitoid rocks.

The mylonite in the Hot Springs area was formed by the microbrecciation of pre-Cambrian crystalline rocks and of Snowbird feldspathic quartzite and arkose. Superficially the rock, despite its original nature, resembles vitreous feldspathic quartzite in appearance; it differs from such quartzite in the presence throughout of well-developed shear planes (which may, however, resemble bedding). In many places it is extremely difficult to distinguish between the mylonite derived from pre-Cambrian granitoid rock and that derived from Snowbird arkose. One useful criterion, however, was afforded by thin sheets of pegmatitic pink feldspar that are not uncommon in the crystalline complex. Because feldspar is one of the minerals most resistant to microbrecciation, these sheets persist as thin and fairly continuous laminae and lenticles of pink feldspar parallel to the shear planes in the mylonite. Recognition of these laminae and lenticles enables one to identify mylonitized pre-Cambrian crystalline rock.

The close resemblance between mylonite and undeformed

[4] Mylonite is defined as a strongly coherent, fine-grained, conspicuously laminated rock formed by extreme microbrecciation and milling of rocks during movement on fault surfaces, with little or no growth of new minerals (see Turner, 1948, p. 201; Waters and Campbell, 1935, pp. 474-476, 478). Petrographic descriptions and photomicrographs of the mylonite in the Hot Springs area appear elsewhere (Oriel, 1950).

feldspathic quartzite led Keith (1904) to map some of the mylonite belts as narrow isoclinal synclines of Snowbird rock in granitic rock. The writer too was deceived by the resemblance during the first half of the field season.

Mylonite is best developed along the bases of the thrust sheets. The distribution of these deformed rocks is indicated by the wavy symbol on figure 2. Closely spaced concentrations of these symbols indicate localities where the mylonite is best developed and thickest. As suggested by figure 2, the writer was able to recognize the presence of the Meadow Fork thrust fault and to trace the western part of the Rector Branch fault largely on the basis of mylonite along the faults.

Other rock behavior near thrusts.—Not all the rocks near the thrust faults are mylonitized; in many places broken, brecciated and somewhat sheared, though unmylonitized, rocks are present. Although rocks are somewhat fractured almost everywhere in the area, the degree of mashing and brecciation is increased with proximity to the faults.

In general, deformed competent beds appear to have yielded by shattering. For example, where the vitreous quartzite beds of the upper part of the Erwin formation lie almost immediately beneath the Mine Ridge thrust surface, they are shat-

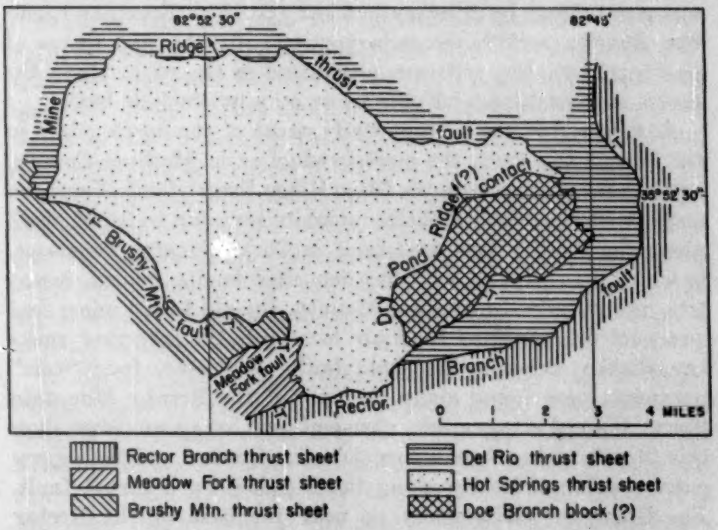


Figure 3. Generalized structural map of the Hot Springs window.

tered and form a fault breccia. Sharply angular fragments of quartzite, ranging in size from minute slivers to blocks several feet long, are embedded in a ferruginous matrix. Such breccia is well exposed about 1200 feet south of U. S. Highway 25-70 and in ledges near the top of the northernmost slope of Lovers Leap Ridge.

Crush breccias have been formed from somewhat less competent beds of arkose and quartzite with some interbedded slate. One good example is exposed immediately below the Rector Branch thrust fault in a road cut on U. S. Highway 25-70, 2200 feet southeast of Walnut Gap.

Incompetent slate, shale, and thinly bedded siltstone are considerably folded, crinkled, and somewhat fractured near the faults, and in places are subphyllitic. The Sandsuck beds exposed in railroad cuts near Runion are an example. That these rocks belong to the Sandsuck formation is clearly shown by exposures of overlying basal Unicoi conglomerate a few hundred feet up the mouth of Big Laurel Creek.

Barite and other mineralization.—In many places throughout the area, fault surfaces and fractures in rocks adjacent to the faults are loci of mineralization (fig. 2). Structural control of the barite mineralization in the Hot Springs district was pointed out by Stuckey in 1942 (p. 107). Specular hematite, fluorite, sericite, secondary quartz, calcite, and traces of pyrite and chalcopyrite are also found in the rocks along the faults in assemblages which may or may not include barite.

As mapped in the present study, most of the barite mines in the Stackhouse area, the most productive in Madison County, are located precisely on the Mine Ridge thrust fault. Ferguson and Jewell (in press) describe mylonite zones up to 9 feet thick along the Brushy Mountain fault in Cocke County, Tennessee, in which mineralizing solutions deposited barite, quartz, hematite, and sericite. In Madison County, the old barite mines and prospect pits north of Bluff are located in the fractured rocks immediately beneath the same fault. Moreover, four "talc" prospects were found along the base of the Brushy Mountain thrust sheet. Petrographic, chemical, and x-ray analyses show this "talc" to be almost pure sericite. Similar sericite aggregates are found locally along the Meadow Fork thrust fault. The Gahagan barite mines lie near the base of the Rector Branch thrust sheet, north-northeast of Walnut Gap. Traces

of specular hematite, barite, sericite, and secondary quartz are also found in many other places along the Rector Branch fault.

Attitudes of thrust faults.—Although the thrust faults can be located accurately within a few feet in many places, the attitudes of the fault surfaces can be measured at only four localities within this area. The well-developed banding in the mylonite, however, suggests a means of determining the dips of the faults. In many places the bedding of the sedimentary rocks and the banding of mylonite below a thrust fault parallel the mylonite banding above it. It appears logical to assume that these well-developed s-planes parallel the thrust surface.

This assumption was tested at the four localities where the thrust surfaces can be observed. At the road cut on U. S. Highway 25-70, 2200 feet southeast of Walnut Gap, the shear planes in both overlying crystalline rocks and underlying sedimentary rocks trend N 40°W and dip 30° NE. The surface of the Rector Branch fault here is not a plane but curves with dips decreasing downward. The average attitude of the fault, however, is about N40°W 25°NE. The Brushy Mountain thrust fault is exposed at Beck Branch and in Turkey Cove, northwest of Bluff. At both localities the fault surface parallels the banding of the silicified mylonite and trends N65°E and dips 45°SE. The most important barite vein exploited in the area lies along the Mine Ridge thrust fault in the Stackhouse mines. Shear planes in the adjacent rocks parallel the vein and the thrust fault which dip 55 degrees to the east.

Thus, the assumption appears to meet the field tests at localities where it can be tested. Nevertheless, the assumption must be considered only tentatively true and is here used as an expedient. Mylonite dips are shown in figure 2. Although 20 degree dips are recorded, the thrust faults dip 40 to 50 degrees in most localities.

In the accompanying structure sections (fig. 4), the major thrust faults are depicted as decreasing in dip downward to a roughly horizontal surface from which the imbricate sheets were upthrust. The representation of the faults beneath the surface profile is, of necessity, hypothetical. The thrusts may dip more steeply even at the depths shown and are probably by no means smooth and planar.

Dry Pond Ridge Contact

Within the Hot Springs window, the contact between the Snowbird formation and the pre-Cambrian crystalline rocks is difficult to interpret. Because it parallels the southeastern flank of Dry Pond Ridge, the name Dry Pond Ridge contact is used here to facilitate discussion of its character. The mapped extent of the contact is shown in figures 2 and 3.

Keith (1904) and Stose and Stose (1947, p. 628) show the

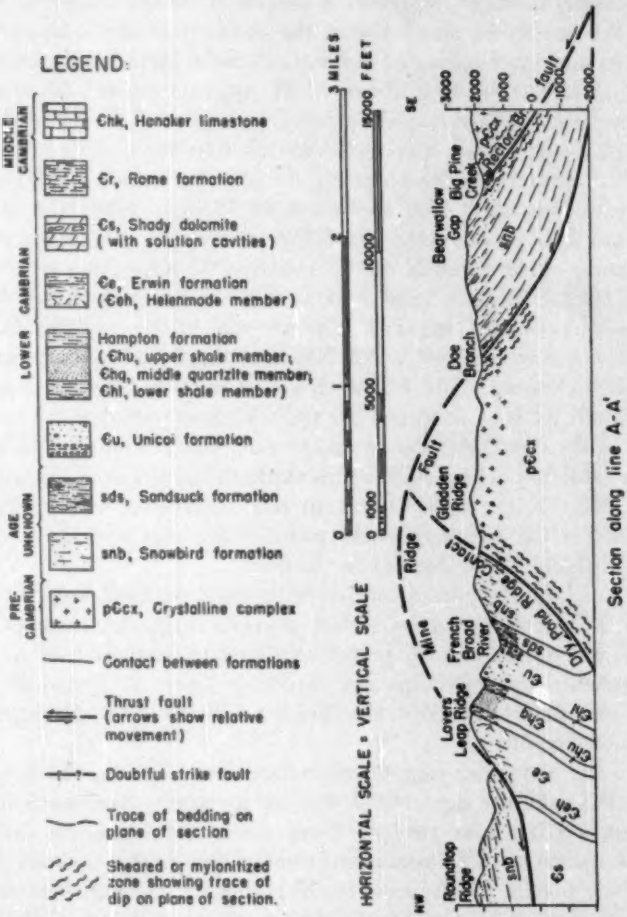
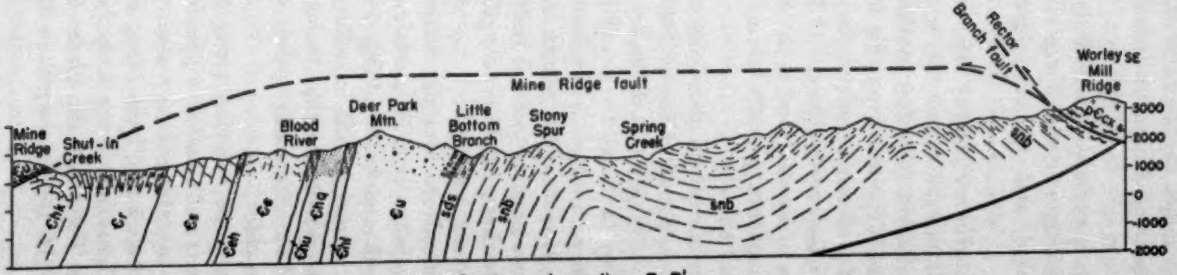
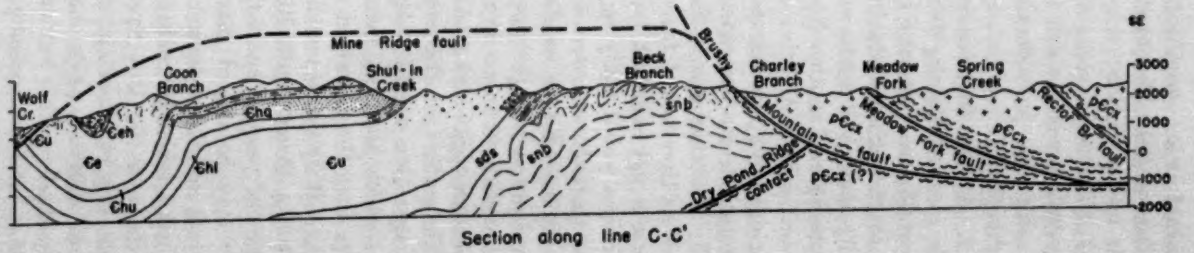


Figure 4. Geologic structure sections across the Hot Springs window. Lines of sections are shown in figure 2.


Mine Ridge fault
Rector Branch Fault
Worley Mill Ridge
SE
Mine Ridge
Shut-In Creek
Blood River
Deer Park Mtn.
Little Bottom Branch
Stony Spur
Spring Creek
3000
2000
1000
0
-1000
-2000
Section along line B-B'


Mine Ridge fault
Brushy
SE
Wolf Cr.
Coon Branch
Shut-In Creek
Beck Branch
Charley Branch
Meadow Fork
Spring Creek
Mountain
Meadow Fork fault
Rector Br. fault
Dry Pond Ridge contact
pCcx (?)
3000
2000
1000
0
-1000
-2000
Section along line C-C'

contact as an unconformity on their maps. On the basis of the following lines of field evidence, however, the writer questions this interpretation:

1. The rocks along the contact, especially the underlying crystalline rocks which have no bedding surfaces along which to slip, are considerably fractured and in many places mylonitized.

2. At least four occurrences of barite are present at or near this contact. Barite, fluorite, and quartz fill fractures in the crystalline rock exposed in the railroad cut 1000 feet southwest of the Stackhouse station. Two barite prospect pits lie in a gully about half a mile northeast of Goforth Cemetery (on the contact) and another lies about a mile west-northwest of Doe Branch school. All other occurrences of barite found in the Hot Springs area are known to be along thrust faults or in the fractured rocks immediately adjacent thereto.

3. The thickness of Snowbird rocks exposed above the contact southwestward from the French Broad River at Stackhouse is not constant (see fig. 2). Although this difference of thickness would easily be explained by faulting, the possibility that the sediments were deposited on an uneven surface cannot be eliminated with available data. However, the lowest beds found are only slightly pebbly and comparable to other beds in the formation; in no way do they resemble the basal conglomerate one might expect on such an uneven surface.

The nature of the contact is difficult to establish because younger beds overlie older. This is always true in unconformities which have not been overturned, and crossbedding indicates that the Snowbird rocks here are right side up. Nevertheless, it by no means eliminates the possibility that the contact is a thrust fault. Billings (1933) cites many known examples of younger rocks thrust upon older.

It may be argued that the evidence of shearing along the contact and the occurrence of barite is not conclusive proof of a thrust fault. In rocks subjected to extraordinary stresses, as those of the Hot Springs area obviously have been, one might expect a certain amount of shearing along a surface of weakness such as an unconformity. The minor slip planes developed along this surface, one might further argue, would localize later mineralizing solutions.

Although this line of reasoning is tenable, the writer never-

theless believes that the Dry Pond Ridge contact is a thrust fault. A third possibility, of course, is that a major thrust fault may have developed along an unconformity.

Despite the lack of conclusive evidence, the Dry Pond Ridge contact is shown as a thrust fault in the structure sections illustrated in figure 4, and the aggregation of sedimentary rocks within the window is here designated the Hot Springs thrust sheet (fig. 3). The mass of crystalline rocks exposed between the traces of the Dry Pond Ridge contact and the southeastern part of the Mine Ridge fault is here designated the Doe Branch block. There is no evidence to indicate whether the Doe Branch block is part of still another thrust sheet or part of the autochthonous rocks underlying the Hot Springs area.

The Hot Springs sheet is overlain on the west, north, east, and southeast by the Del Rio thrust sheet, and on the southwest and south by the Brushy Mountain, Meadow Fork, and Rector Branch thrust sheets.

Other Faults

In addition to the thrusts, four minor faults were mapped in the rocks within the window. The most extensive of these faults is the one trending northwest from the southwestern end of Deer Park Mountain across Shut-in Creek to Dry Branch. Although the fault surface itself was not seen in any outcrop, the writer believes the fault well substantiated by the abrupt offset of the Hampton formation (especially the easily traceable middle quartzite member) and of the upper ledge-forming quartzites of the Erwin formation. The apparent relative displacement of the southwest block is to the northwest. Truncation of formational boundaries occurs along a comparatively straight line despite moderately rugged topography, thereby suggesting that the fault has a steep dip.

Another fault with a smaller relative displacement in the same direction occurs just north of Stony Spur and west of Spring Creek. The presence of the fault is easily recognized where the lower Unicoi conglomerate beds on the northeast side of the fault are in contact with Sandsuck slate southwest of it.

The other two faults strike northeast and the apparent relative displacements of the southeastern blocks are to the northeast. The fault south of Hot Springs and just east of Spring

Creek was recognized by the slight offset of the basal Unicoi conglomerate beds across their strike. The evidence for the fault southeast of Hot Springs and east of French Broad River is the abrupt change from Hampton middle quartzite beds to upper Hampton shale beds along a contact normal to the strike of the bedding.

Although all four of the faults described here have steep dips, the lack of other data makes it difficult to classify them except as dip faults. The fault first described, however, offsets nearly vertical beds for a considerable distance and its displacement is probably at least partly, if not wholly, strike-slip.

The writer was unable to find any evidence to support the cross fault inferred by Stose and Stose (1947, pp. 641-644) in their discussion of the origin of the hot springs.

Folds

The magnitude of the folds developed in the strata of the Hot Springs area ranges from microscopic to several miles. Inasmuch as the largest folds are quite evident on the geologic map (fig. 2) and in the structure sections, the discussion which follows deals largely with the others.

The rocks of the Honaker formation are horizontal, or nearly so, where exposed in Mine Hollow, whereas the underlying rocks, as far southward and down the section as the Unicoi formation, dip steeply to the north. Efforts to determine whether the Honaker limestone beds are right side up proved unsuccessful. If the beds are right side up, then the outcropping Honaker beds lie along the axis of a normal syncline. If the beds are upside down, however, then the fold in which they appear is an "anticline", as shown in figure 4, section B-B′. Both are possible. However, the writer believes the "anticline" interpretation the more likely because one would expct vertical beds to be affected by the northward drag near the over-riding thrust fault.

The word anticline is placed in quotes above because in this interpretation the structure is really an upside-down syncline. A normal anticline is a fold with older rocks toward the center of curvature. In the fold shown in figure 4, section B-B′, one passes into younger rocks in going towards the center of curvature.

Comparable folds may be present in other localities in the

area. The structure of upper Erwin strata exposed in the long road cut near the Hot Springs school, on U. S. Highway 25-70, is described by Stose and Stose (1947, pp. 630, 641-642) as anticlinal. Although the beds are arched up, the possibility that the sequence is overturned is suggested by the downward pinching out of quartzite beds which somewhat resemble coarse upside-down crossbedding. Moreover, in the poorly exposed but critical portion of the outcrop, on the north side of the large fold, there is a suggestion that the strata swing up again. South of the axis of the fold, all strata dip steeply to the north and are right side up. Nowhere, south of the U. S. Highway bridge across Spring Creek, are dolomitic rocks of the Shady formation present in the fairly continuous exposures along the bluffs of Spring Creek, as one would expect if the fold were a normal anticline.

Folds in the Shady dolomite are exposed along the east bank of the French Broad River opposite the town of Hot Springs. The folds measure tens of feet from crest to crest and the attitudes of their limbs are approximately N-S 30-35°E and E-W 50-70°N. The axial planes of these folds thus dip to the southeast. The proximity of these folds to the Mine Ridge thrust fault suggests that they were formed as a result of drag along that fault surface.

In the area mapped, rocks of the Rome formation are highly contorted. Folds are intricate and range in magnitude from a few inches to at least tens of feet. The multifarious trends of the axial planes of the folds make it difficult to determine the relation of these folds to the broader structural features of the area.

The writer found no structural evidence supporting Keith's (1904) interpretation of a syncline which repeats the Shady dolomite in a belt along the northernmost part of the window. Stratigraphic observations, on the other hand, indicate the presence of Honaker limestone within the window and show that the Rome formation includes interbedded Shady-like dolomite beds which Keith may have assigned to the Shady formation, along with the Honaker, where the enclosing shale was not exposed.

Cleavage

Cleavage is present in the fine- and very fine-grained clastic rocks beneath the Shady dolomite in many, but not all, local-

ities in the area. The cleavage appears to be best developed in the rocks of the Sandsuck formation. The nature of the cleavage differs with the lithology. In the predominantly argillaceous beds, cleavage planes are closely spaced and parallel the planes of the preferentially oriented sericite plates. In rocks with larger silt and sand fractions, cleavage planes are not so numerous and somewhat resemble closely spaced joints.

In general, cleavage planes strike to the northeast and dip about 50 degrees to the southeast. Local exceptions include rocks in which cleavage parallels bedding which does not dip to the southeast and rocks in which cleavage intersects bedding and dips to the east. Where bedding does dip to the southeast, cleavage closely parallels the bedding.

In some shales and slates, cleavage surfaces are folded and crinkled. The crinkles appear to have developed by minor slip along the bedding (which in most places forms a steep angle with the cleavage) after the cleavage planes formed in the rock.

Relative Direction of Thrust Movement

The following points support the conclusion that the relative movement of each thrust sheet in the Hot Springs area was approximately from the southeast to the northwest with respect to underlying rocks:

1. Most of the thrust faults dip to the south, southeast, and east. One notable exception is the Dry Pond Ridge contact, if it is a thrust fault, which dips to the northwest and west, but it is possible that the root zone for this fault is concealed beneath the southeastern part of the Del Rio sheet. The other exception is the northern part of the Mine Ridge thrust fault which dips to the north and northwest, but the root zone is exposed in the southeastern part of the area mapped. An attempt is made in later pages to explain the northward dips of these faults.
2. In general, cleavage planes dip to the southeast and east.
3. Axial planes of drag folds in the Shady dolomite, just below the Mine Ridge thrust fault, dip to the southeast.
4. Almost all the thrust sheets described northwest of the Piedmont in the Southern Appalachians are believed to have moved relatively to the northwest (*e.g.* Butts, Stose, and Jonas, 1932, p. 7; Butts, 1940, pp. 440-464; King *et al.*, 1944, p. 11). From these and other structural features, it has been concluded

(Chamberlin, 1928, p. 90) that the region was subjected to stresses which might be described as a couple in vertical planes trending NW-SE. Forces near the surface were directed northwestward while deeper in the crust they were directed southeastward. According to this interpretation, it is not likely that any of the major thrust sheets in the Hot Springs area moved relatively to the southeast.

Therefore, it is concluded from both field observations and regional considerations that the movement of the thrust sheets in the Hot Springs area was northwestward relative to the rocks below them.

It does not appear possible, on the basis of existing data, to determine the magnitude of the fault displacements, nor their southeastward extent. However, it is likely that the minimum figure would be a few, or possibly tens of miles.

Summary of Structural Data

From the foregoing discussion, it is evident that the Hot Springs window is not a simple window which has formed by erosional penetration through a single thrust sheet to expose the underlying, overridden rocks. The frame is made up not of one, but of four different thrust sheets. Moreover, the existing data which suggest that the Dry Pond Ridge contact is a thrust fault also indicate that a smaller window, framed by two different thrust sheets, may be exposed within the Hot Springs window.

Noteworthy features of the window which will be further discussed below may be summarized as follows:

1. The window is framed by more than one thrust fault.
2. Both the axis of elongation of the Hot Springs window and the general bedding of the rocks within the window strike across the regional structure of the Southern Appalachians. The general bedding and the axis of elongation of the window strike about east-west whereas regional structures strike predominantly NE-SW.
3. Fault surfaces around the window dip steeply (40° to 50°) away from the window.
4. The fault surfaces framing the window converge downward at a considerable angle and merge to form the major movement zone (fig. 4).
5. Mylonite is present along portions of each of the thrust faults.

6. The Hot Springs thrust sheet within the window may be arched over the granitoid rocks of the Doe Branch block. This cannot be proved inasmuch as the rocks beneath the Del Rio sheet and southeast of Doe Branch are concealed. However, trends of the Dry Pond Ridge contact and the overlying sedimentary rocks suggest that a southwest plunging anticline may be present within the window.

7. Reference to a regional geologic map (see Geologic Map of the U. S., 1932, or Tectonic Map of the U. S., 1944) indicates that the Hot Springs window lies along the projected axis of the Mountain City window in northeastern Tennessee. Except for the Grandfather Mountain window, the rocks in the Hot Springs and Mountain City windows constitute the southeasternmost exposures of known Cambrian formations in this segment of the Appalachians.

COMPARISON WITH OTHER COMPLEX WINDOWS

Before attempting to trace the sequence of tectonic events which led to the formation of the Hot Springs structure, the writer found it instructive to examine a few other windows with multiple faulting. Such complex windows are not unknown although comparatively few have been described.

Southern Appalachian Examples

Rose Hill district.—The Rose Hill district in Lee County, Virginia (Miller and Fuller, 1947), is notable both for the complex structural relations and for the oil production obtained from the Cambrian and Ordovician carbonate rocks beneath the Pine Mountain thrust fault. Thirteen windows, some of which are small windows within larger ones, are present within five miles almost along the axis of the Powell Valley anticline. "In much of the fenster area the Pine Mountain overthrust consists of a fault zone having well defined upper and lower planes of movement, between which there are slices of folded and faulted rocks" (*op. cit.*). These upper and lower movement planes are roughly parallel. Some of the windows, *e.g.*, the Hamblin Branch and Possum Hollow fensters, are framed by only one plane of movement. Others, *e.g.*, the Dean, Fourmile, and Sugarcamp fensters, are parts of the stationary block framed by both the upper and lower branches of the Pine Mountain thrust, *i.e.* each is bounded on one side by rocks of the Cumberland thrust block and on the other by the fault slices.

Moreover, these latter windows lie within a larger window framed only by the Cumberland thrust block.

Draper Mountain structure.—The Draper Mountain structure in southeast Virginia is described by Cooper (1939, p. 71) as a "breached anticlinal compound fenster" which has been "formed by the crenulation of two thrust sheets and subsequent erosion through both of them". The breached window is almost completely framed by the Pulaski and the Max Meadows thrust faults which roughly parallel each other. It is possible that the Max Meadow fault is merely a lower plane of movement and forms, with the Pulaski fault, a single zone of movement, comparable to the Pine Mountain fault zone.

Roanoke district.—Unlike the aforementioned windows, the Round Hill and the Coyners-Mills fensters, as well as the Goose Creek breached fenster, in the Roanoke area, Virginia ((Woodward, 1932, plate 1, figs. 4, 5, pp. 80-90; Butts, 1932, map and sections J-J′), have each been mapped as windows framed by a single thrust sheet. The first two lie beneath the Pulaski thrust sheet, whereas the Goose Creek breached fenster is almost completely bounded by the Blue Ridge thrust sheet. Despite the presence of simple windows, the area is mentioned here because of the multiple faulting. The Blue Ridge thrust sheet lies above the Pulaski sheet, which in turn overlies the Saltville fault block. Although the Blue Ridge and Pulaski faults do not unite at the surface, they may at depth; indeed, they are so shown in the structure sections by Woodward and by Butts. Thus, it is probable that the Roanoke structure compares with the Hot Springs structure in the downward merging of the faults underlying the imbricate thrust sheets.

Mountain City window.—The Mountain City window in northeast Tennessee (King *et al.*, 1944, pp. 11-13, fig. 2, plate 1) is comparable to the Hot Springs window in several ways, although it is considerably larger and more complex. The window is framed on the northwest by the Iron Mountain thrust fault and on the southeast by the Stone Mountain thrust fault. The structure is complicated, however, by the cross faults which offset the framing thrusts and by the many thrust faults within the window. The Iron Mountain and Stone Mountain faults unite at the surface on the south and northeast sides of the window and thus converge downward and merge to the south-

east (*op. cit.*, plate 1, sections D-D′ and E-E′). Moreover, within the Mountain City window lies another, the Limestone Cove window. This window too is immediately overlain by two different thrust sheets: the sheet overlying the Limestone Cove thrust fault on the northwest and the Stone Mountain sheet on the southeast. The faults beneath these thrust sheets also converge to the southeast and merge, albeit upward due to folding of the thrusts. Thus, in the Mountain City and Hot Springs windows, as well as in the Roanoke area, the convergence of the faults is in the direction from which the thrust sheets are believed to have moved.

Engadine Window

The Hot Springs and Mountain City windows are closely analogous in form to the classic Engadine window of the Central Alps. For this reason, a somewhat fuller discussion of that window is presented here. Much of the material which follows was drawn from a detailed study by Hammer (1915) and from summaries by Sander (1921, pp. 193-195, 198, 213-214) and Klebelsberg (1935, pp. 139-146).[5]

Geology of Engadine window.—The Engadine window is subovate in form and trends NE-SW in a region (only part of which is represented in figure 5) where E-W trending structures predominate. The rocks within the window are named the Bündner Schiefer and include calcareous schist, mica schist, chlorite schist, and phyllite of Mesozoic age. The Bündner Schiefer forms a large symmetrical anticline whose limbs dip steeply and uniformly to the northwest and southeast; the axis of the anticline trends NE-SW and passes almost through the middle of the window.

The frame on the north side of the window consists of the Silvretta mass of Early Paleozoic gneisses. The Silvretta mass is a rootless thrust sheet which has moved from the southeast to the northwest. The Oetztal mass frames the window on the southeast and is made up of rocks similar in age and lithology to those in the Silvretta. The older rocks of both the Oetztal and Silvretta masses rest upon the Bündner Schiefer above thrust surfaces which dip steeply (40° or more) away from the

[5] Less comprehensive summaries of the geology of the Engadine window have been published in the English language by Collet (1927, pp. 27, 215, 230-231) and by Heritsch and Boswell (1929, pp. 27-28, 66, 70-76, 102-105, 153-154).

window. Only in a few localities do the thrust faults dip more gently. Mylonite is fairly well developed along the base of both thrust sheets. Northeast and southwest of the window, the Oetztal mass has been thrust over the Silvretta. Thus, the Silvretta gneiss has been thrust over the Bündner Schiefer and both have been overridden by the Oetztal gneiss. As in some of the aforementioned structures, the fault beneath the Oetztal mass diverges upward to the northwest from the fault beneath the Silvretta, *i.e.* the faults converge downward and merge in the direction from which the thrust sheets are believed to have moved.

Definition of eyelid window.—The Engadine fenster, then, is not a simple window but one framed by two different thrust surfaces. For this reason, Sander (1921, p. 193) proposed the term shear-window (*Scherenfenster*) and defined it as a window whose rocks are directly overlain by more than one thrust sheet. Sander emphasizes that the term is descriptive and not genetic. Shear-window is rather an unfortunate term, however, inas-

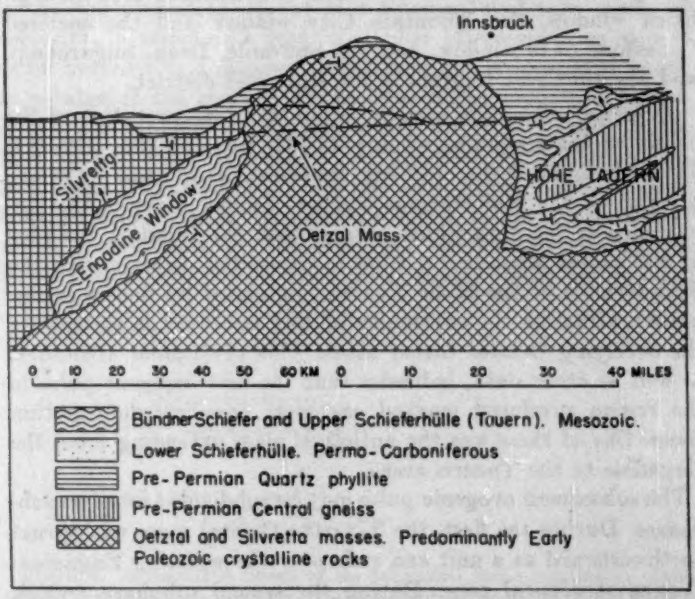


Figure 5. Generalized structural map of the western Tyrol, showing the Engadine window and part of the Tauern. (Modified from Sander, 1921.)

much as shearing is implicit in the formation of any and all windows.

Another term is suggested by Sander's description (1921, p. 212) of the ends of the window. He points out that the framing thrust sheets overlap, as in an iris diaphragm. A more restricted translation of the components forming the word *Irisblende* could be interpreted as "eyelid". The writer has heard several geologists use the expression "eyelid window" but he has been unable to find any reference to it in the geological literature. Therefore, it is here proposed that *eyelid window* be defined as Sander defined *Scherenfenster* and that the latter term be abandoned. The proposed name has the merit of suggesting the analogous form of two eyelids (thrusts) over the enclosed eye (rocks within the window). The analogy may be criticized, however, because it suggests movement from opposite sides of the window, whereas known examples are believed to have formed by thrusting in one direction.

American examples of eyelid windows, as described above, are the Hot Springs window and the enclosed Doe Branch block window, the Mountain City window and the enclosed Limestone Cove window, and the Fourmile, Dean, Sugarcamp, and Martin Creek fensters in the Rose Hill district.

Origin of Engadine window.—Sander's interpretation of the origin of the Engadine window (1921, pp. 193-196, 212-214, 219), though not the only one published,[6] is in some ways applicable to the Hot Springs window. Sander stresses the significance of the data which suggest that the Bündner Schiefer of the Lower Engadine window is connected directly with the Schieferhülle of the Hohe Tauern (indicated by dashes in figure 5), although the continuity of this structure is masked by the overlying Oetztal thrust sheet. This continuous structure, as well as other data, indicates that the first orogenic pulse in the region produced marked east-west trending deformation zones. One of these was the anticlinal mass extending from the Engadine to the Tauern areas.

The subsequent orogenic pulse may be subdivided into two subphases. During the first, the Silvretta-Oetztal mass was thrust northwestward as a unit and rode over the resisting Engadine-Tauern structural zone. During the second subphase (which

[6] Two more recent, but conflicting, interpretations are those by Staub (1937, pp. 18-108) and Krasser (1940, pp. 181-185).

was probably continuous with the first in time) the frictional resistance along the major thrust surface, beneath the Silvretta-Oetztal mass, became too great and the Oetztal subsidiary thrust sheet split off and rode over the base of the main mass. The sliver left beneath the Oetztal mass is the Silvretta thrust sheet. The tangential force couple that emplaced the Oetztal thrust sheet acted in a NW-SE direction in the Engadine area (fig. 6), producing a counter-clockwise rotation of the underlying Bündner Schiefer mass so that it no longer parallels the closely related east-west zones. Sander and other Austrian geologists emphasize that the present imbricate structure is the result of the formation of subsidiary thrust sheets when the frictional resistance became too great along the main movement horizon. They deny that it formed by the piling up of far-travelled thrust masses developed from recumbent folds, as the Nappists believe.

POSSIBLE TECTONIC EVENTS IN HOT SPRINGS AREA

Any attempt to describe the sequence of tectonic events which led to the formation of the Hot Springs structure must, of necessity, be based on a certain number of assumptions as well as on available field observations. There are serious gaps in our knowledge of the geology of the Blue Ridge mountain region. What, for example, is the relation of the thrust faults in the Hot Springs area to those of northeastern Tennessee and the Mountain City window? Does the Hot Springs thrust sheet form an arch over the Doe Branch block, the southern limb of which is concealed by the overlying thrust sheets? Are the Brushy Mountain and Meadow Fork thrust sheets not exposed

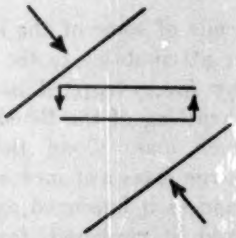


Figure 6. A simplified sketch of the tangential forces applied to earlier formed east-west structural zone (as indicated by arrow in figure 5). The stresses in the NW-SE direction resulted in the counter-clockwise rotation of the structures in the Engadine window. (After Sander, 1921, p. 194.)

outside the eastern half of the window because they have been overridden by the Rector Branch sheet, or do their eastern limits mark the line of a partially concealed tear fault? The answers to these and comparable questions would probably establish or disprove the validity of the statements that follow.

It is the writer's belief that the Hot Springs window was formed in much the same way as the Engadine window, according to Sander's interpretation. The steps in its formation are conceived to be somewhat as follows: If the Dry Pond Ridge contact is a major thrust fault, an early tectonic event was the thrusting of the Hot Springs sheet over the Doe Branch crystalline rocks. After this movement, forces applied in a NW-SE direction compressed the rocks of the Doe Branch block and the Hot Springs sheet to form a large NE-SW trending anticline with steeply dipping limbs. This anticline may have been formed at the same time as, and may have been continuous with, comparable structures in the Mountain City window vicinity.

The first phase was followed by the relative northwestward movement of the Del Rio sheet along the Mine Ridge thrust fault and across the arched Hot Springs sheet, truncating the latter and partially severing it from its root. The absence of the Del Rio sheet beneath overlying thrust faults in the southwestern part of the area and the decrease in the width of the sheet as exposed southwestward from the French Broad River suggest that the Del Rio sheet may wedge out southward. Thus, the remaining thrust sheets may be interpreted as imbricate blocks of the same mass or as parts of another major thrust mass which bevelled off the Del Rio sheet. The former of these possibilities is illustrated in the accompanying structure sections (fig. 4).

The incongruous east-west trends of some of the structures in the Hot Springs area may be attributable to the clockwise rotation of earlier structures by forces applied in a north-northwest direction during the thrusting of the Brushy Mountain-Meadow Fork-Rector Branch mass along the Brushy Mountain thrust fault. As this large mass met increasing frictional resistance along its base and as it impinged against the structural front already developed, it ruptured, forming the Meadow Fork and the west end of the Rector Branch thrust fault.

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[American Journal of Science, Vol. 249, January, 1951, Pp. 31-46]

EARLY CRETACEOUS MAMMALS FROM NORTHERN TEXAS

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ABSTRACT. A new triconodont, *Astroconodon denisoni*, based on lower jaws, and an undetermined mammal, represented by an incomplete humerus, are described from a locality in the upper part of the Trinity Sand, near Forestburg, Montague County, Texas. This is the first locality yielding Early Cretaceous mammals found in America and the second in the world. *Astroconodon* is more advanced than Late Jurassic triconodontines, particularly in the tongue and groove interlocking of the lower molars. The angle between the long axes of the proximal and distal ends of the undetermined humerus is considerably greater than in the Theria. The specimens were found, accompanied by numerous fragmentary remains of fishes, frogs and reptiles, in small pockets, believed to have been laid down under near-shore conditions. The occurrence is of early Albian age.

ONE of the pleasant features of paleontology is that so many of its discoveries are accidental. The exciting find here reported is of this sort, a totally unexpected by-product of a search for turtles that was planned with no expectation of finding a new and promising locality for Mesozoic mammals.

Prior to attending the Annual Meeting of The Geological Society of America at El Paso, Texas, in November of 1949, Dr. Rainer Zangerl and Dr. Robert H. Denison undertook a brief reconnaissance of some exposures in the Early Cretaceous Trinity Sand near the town of Forestburg in Montague (pronounced Mon-taig) County, Texas. Dr. Zangerl's interest had been aroused in this area as a possible source of Early Cretaceous turtles by Mr. Glen L. Evans, Assistant Director of the Texas Memorial Museum. Mr. Evans, in turn, had been informed of the possibilities of the region by Mr. Louis H. Bridwell, lepidopterist and entomological collector, of Forestburg. Drs. Zangerl and Denison, guided by Mr. Bridwell, were examining some exposures about two and a half miles from the town when they both noticed a concentration of small bone fragments on the surface. After a few minutes' search, Dr. Denison picked up the fragmentary lower jaw of a triconodont —the first Early Cretaceous mammal to be found in the New World. (The only other known Early Cretaceous mammalian occurrence is in the English Wealden, a formation which has yielded a few isolated teeth, all either multituberculate or indeterminate.) This specimen was brought on to El Paso, ex-

amined and determined under a binocular microscope and the discovery briefly reported on before The Society of Vertebrate Paleontology.

The incomplete humerus described below was also discovered during this first visit, but its mammalian affinities were not realized until it was re-examined in the laboratory.

Following the meetings, Drs. Zangerl and Denison, accompanied by Dr. and Mrs. George Gaylord Simpson and myself, returned to Forestburg for a stay of three days. During this time, a number of bone-bearing pockets were found, and Dr. Denison further distinguished himself by finding a second—and better—triconodont jaw. In addition to the mammals, the pockets yielded fragmentary remains of many vertebrates: small sharks, pycnodonts and various undetermined fishes; frogs; turtles; small lizards; crocodiles; carnosaur and ornithopod dinosaurs; and an occasional fragment of a pterosaur. Outside the pockets, fragmentary remains of ornithopod dinosaurs were encountered. Before leaving, some sacks of matrix were taken from the two pockets that had yielded the triconodont jaws. These were washed and sifted in the laboratory and the "concentrate" carefully sorted over. This patient toil resulted in the finding of numerous additional specimens of the forms just mentioned above.

A brief, preliminary note on the discovery by Zangerl and Denison, now in press, will have appeared by the time this article is in print. A program of co-operative field work in the Trinity Sand of northern Texas by the Texas Memorial Museum and the Chicago Natural History Museum will be carried out during the present year.

To all those persons mentioned above, to Mr. Vergil Greenwood, owner of the land on which the specimens were found, and to Mr. Douglas Tibbetts, who made the drawings, I wish to express my sincere thanks.

Order Triconodonta

Family Triconodontidae

Subfamily Triconodontinae

Astroconodon[1] gen. nov.

[1] ἀστήρ a star (lone) + ἄξη point + ὀδούς tooth; named for the state in which it was found and with analogy to related forms. The trivial name is in honor of Dr. Robert H. Denison, discoverer of the specimens.

Type species: *A. denisoni*[1] sp. nov.

Distribution: Early Cretaceous, Texas.

Diagnosis: Differing from the Late Jurassic *Triconodon*, *Priacodon* and *Trioracodon* as follows: posterior cingulum cusp of P_4 larger, pointed and projecting posteriorly; posterior cingulum cusp of lower molars larger and higher; anterior cingulum cusp of lower molars completely incorporated in anterior groove; this structure better developed than in Jurassic forms, involving entire anterior face of tooth and extending down into anterior root; necks of cheek teeth cement covered, cement on molars forming posterior spur projecting into anterior groove of tooth following. Lower molars completely interlocking. Internal mandibular groove very faint.

Astroconodon denisoni sp. nov.

Type: CNHM no. PM 542, portion of right ramus with roots of penultimate premolar, incomplete last premolar, M_{1-2}, anterior root of M_3; a fully adult individual with well worn teeth.

Hypodigm: Type and CNHM no. PM 541, portion of left ramus with roots of last premolar, incomplete M_1, M_2, incomplete anterior alveolus of M_3; an old individual with deeply worn, somewhat abraded teeth.

Horizon: Exact position in rock and time uncertain, but occuring within the upper 100 feet of the Trinity Sand, and hence somewhere in the early Albian; discussion below.

Locality: Approximately 2½ miles southwest of Forestburg, Montague County, Texas.

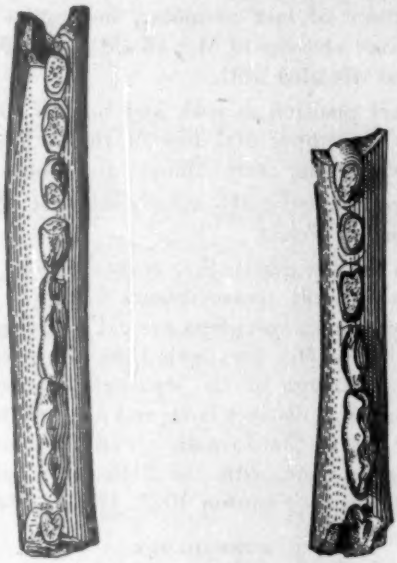
Diagnosis: As for the genus. Size comparable to that of most Late Jurassic species; measurements below.

Since the two known specimens are old individuals with worn teeth, knowledge of the few teeth that are preserved is far from complete. Enough of the structure is shown to reveal that *Astroconodon* is distinct from and more advanced in some respects than any of the Jurassic species, but not enough to permit full comparison with the little-worn teeth that are known in these forms (Simpson 1925, 1928a, 1929).

MORPHOLOGY

The last premolar of the type is unfortunately broken through the middle, and the central main cusp (cusp b of Simpson 1925, protoconid of Butler 1939b) is almost entirely

missing; no idea of the height of this tooth relative to the molars can be gained. It is slightly longer than M_1. There is a prominent anterior cingulum cusp (anterior cusp of Butler, 1939b), nearly worn away in the specimen, that stood free of the anterior slope of the main cusp and appears to have been larger than in any of the Jurassic forms. The posterior main cusp (posterior accessory cusp of author's, cusp C of Simpson, 1925) is heavily worn and was undoubtedly more distinct in earlier stages of wear. In its present state, it merely forms a convexity on the internal side of the posterior slope of the main cusp and is barely visible externally. The posterior cingulum cusp (posterior cusp of Butler, 1939b) is well developed and sharply pointed; it projects straight posteriorly well beyond the posterior root and toward the deep groove in the anterior face of M_1. A well defined and quite broad cingulum runs forward from the apex of the posterior cingulum cusp; in all probability, this reached to the anterior cingulum cusp, although breakage prevents certainty on this point. There is no trace of an external cingulum. The roots



A. B.

Fig. 1. *Astroconodon denisoni* gen. et sp. nov. Crown views of: A, type, CNHM no. PM 542; B, referred specimen, CNHM no. PM 541. X6.

are oval in section and somewhat longer than in the molars; those of the penultimate premolar indicate that this tooth was very nearly equal in size to the last of the series.

At least three molars were present, of which M_3 is represented by only the incomplete anterior root in the type and by part of the anterior alveolus in PM 541. As in the Jurassic triconodontines, M_{1-2} are similar in structure, and M_2 is longer than M_1. M_2 is considerably wider than its predecessor. The labial margins of both teeth are in line, up until a very advanced stage of wear is reached (M_1 of PM 541), but the lingual margin of M_1 is inset relative to that of M_2. Both teeth are essentially even in width across the main cusps, *i.e.* for most of their lengths, but, due to the prominence of the posterior cingulum cusp, taper posteriorly to a greater extent than would appear to be the case in the Jurassic forms.

Crown structure is seen most clearly in M_2 of the type, the least worn tooth available. In common with the Jurassic forms, there are three main cusps arranged in an antero-posterior line; the central cusp may have been slightly larger and higher than the anterior (anterior accessory cusp of authors; cusp a of Simpson, 1925; paraconid of Butler, 1939b) and posterior, but the degree of wear prevents certainty on this point. Between these cusps, on the lingual side and no doubt originally on the labial, are well defined grooves that extend to and taper toward the base of the crown. The notch separating the central and posterior cusps is, in the present state of wear at least, rather deeper than that between the central and anterior. The posterior cingulum cusp is unfortunately damaged on this tooth and badly worn on the others, but it was evidently larger, perhaps higher and certainly produced farther posteriorly than in the Jurassic genera. The anterior cingulum cusp is incorporated into the anterior vertical groove. This remarkable structure, foreshadowed in Jurassic triconodontines (Simpson 1925, p. 150; 1928a, p. 81), is a deep channel occupying most of the anterior face of the tooth for its entire height. Beneath, it continues into the root as a V-shaped slot that received a sharp crest projecting from the anterior wall of the alveolus. The sides of the crown portion of the groove are formed by the antero-external part of the anterior main cusp labially, and by the anterior edge of this cusp, plus the anterior cingulum cusp, lingually. The reasons

for believing that the anterior cingulum cusp participated here are that the lingual side of the groove is larger and extends farther anteriorly than the labial and that the internal cingulum runs forward to a point where a cingulum cusp is known to occur in some Jurassic triconodontines (Simpson 1925, p. 150; Butler 1939b, fig. 8d). The posterior end of each molar crown forms a tongue that fits snugly into the groove of the tooth following and this is also true of the extra-alveolar part of the posterior root, as pointed out below. The entire molar series thus forms a single, interlocked, uninterrupted cutting edge. Although the anterior groove of M_1 is as perfectly formed as that of M_2, the posterior portion of the last premolar is by no means completely adjusted to it, and there is no indication that this lack of good fit has been brought about by post-mortem displacement. This is a point of some general interest, and will be discussed further in the succeeding section.

The labial face of the tooth under description is heavily worn, nearly plane anteriorly and slightly concave behind the central cusp, where it was in occlusion with the anterior part of M^2. The cutting edge, which of course follows the notches between the main cusps, is decidedly sinuous; as shown by the other available teeth, this edge becomes straighter as wear proceeds and the notches are progressively eliminated. Due to the type of occlusion in triconodonts, the edge is maintained throughout practically the entire life of the tooth. Small pits on the sides of the main cusps on this face mark where wear has reached the pulp cavity with resulting deposition of secondary dentine (fig. 2c).

There is no trace of an external cingulum. The internal cingulum is rounded rather than sharp, fairly straight for most of its length but turning up abruptly toward the posterior end of the tooth to terminate at the cutting edge in the posterior cingulum cusp.

The roots of the molars, particularly the anterior ones of each tooth, differ considerably from those of the premolars. The anterior is the shorter of the two and narrows to a blunt edge posteriorly; the anterior groove described above gives to the section a shape somewhat resembling that of a conventional heart. The posterior root is rounded anteriorly and tapers to a sharper edge posteriorly.

The extra-alveolar portion of the roots—the neck of the

tooth—is heavily invested by cement in the molars, and to some extent in the premolars also. This is particularly well shown in the type, less so in the more abraded referred specimen. The

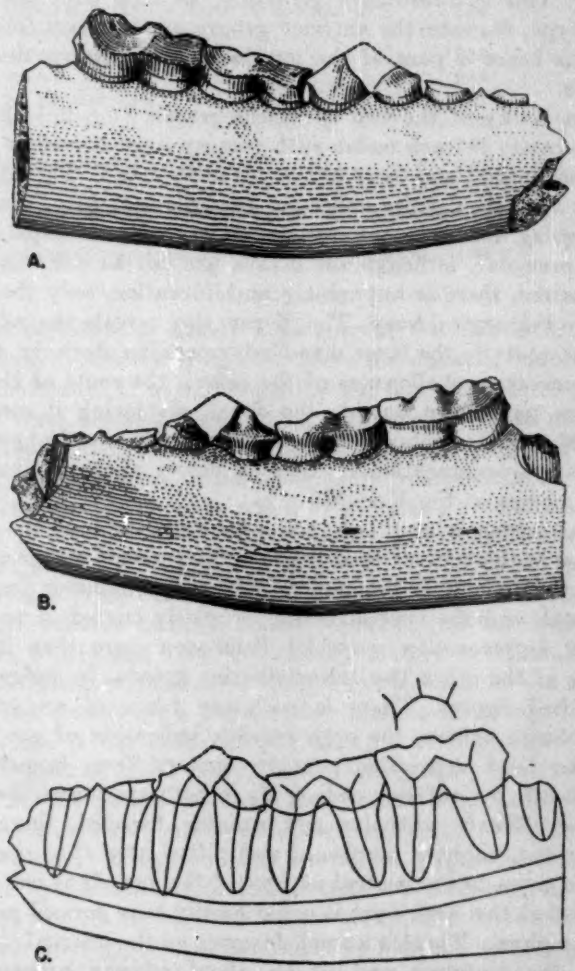


Fig. 2. *Astroconodon denisoni* gen. et sp. nov. A and B, external and internal views of type, CNHM no. PM 542. C, diagram based on X-ray of type showing roots, pulp cavities and mandibular canal; inset shows upper part of pulp cavity of M_2 of referred specimen. CNHM no. 541. Offsets in the roots at the alveolar border in C are due to the thickness of the extra-alveolar cement covering. X6.

cement on the molars fills the space between the roots, forms a shelf on either side beneath the enamel line and curves abruptly up posteriorly, where it forms a blunt, posteriorly projecting spur. This extraordinary structure, seen on both molars of the type, fits into the anterior groove of the tooth following, and is hence a part of the interlocking mechanism described above.

An X-ray of the two specimens reveals (fig. 2c) that the pulp cavity in each molar root goes up approximately to the enamel border, and there bifurcates to form a Y. The adjacent arms of each Y unite under the central main cusp, the other two going under the anterior and posterior main cusps. In the last premolar, although the details are not as clear as could be desired, there is apparently no bifurcation, only the union under the central cusp. The X-ray also reveals the relations of the roots to the large mandibular canal; anteriorly, due to the increasing shallowness of the ramus, the roots of the premolars pass down outside the canal, displacing it medially. In PM 541, the roots are somewhat longer than in the type, and there is a prolongation of the pulp cavity above the junction of the arms of the Y in M_2.

The ramus is stout compared to the widths of the teeth, and is, in the type at least, rather wider across the premolars than the molars. It becomes progressively shallower from M_1 forward, and the ventral border is gently curved, a point in which *Astroconodon* resembles *Priacodon* more than it does either of the other two triconodontine genera, to judge from published figures. There is no clear trace of an internal mandibular groove, the only possible indication of one being a wide, faint depression running forward from beneath M_3 to beneath the anterior end of M_2 in both specimens. Beneath the penultimate premolar and running forward, there is a rather flat, slightly roughened and pitted area that does not extend down to the ventral margin of the jaw. It is not at all clear what this area was; it could hardly have formed part of the symphysis. There is a small foramen on the internal side beneath the posterior root of M_1 about midway between the alveolar border and ventral margin, and another, slightly lower, beneath the posterior root of M_2; the anterior foramen of this pair, at least, is absent in the referred specimen. No mental foramen is present in either jaw; as in *Triconodon* and

Trioracodon, this opening, or openings, was more anterior in position.

Measurements in mm.

	PM 542, type	*PM 541*
L. penultimate premolar roots at alveolar level	2.58	——
L. last premolar crown	2.71	——
L. last premolar roots at alveolar level	2.59	2.28
L. M_1 crown	2.44	——
L. M_1 at alveolar level	2.37	2.35
W. M_1	.59	——
L. M_2 crown	2.80	2.41
L. M_2 roots at alveolar level	2.58	2.24
W. M_2	.78	.70
Depth of ramus below anterior end of last premolar	3.26	2.92
Depth below posterior end of M_2	3.80	3.74
Width of ramus across anterior end of last premolar	1.84	1.54

DISCUSSION

Astroconodon is clearly a member of the Triconodontinae, and equally clearly is distinct from the Jurassic members of the subfamily. It is unfortunately not now possible to go beyond these simple statements in a discussion of its affinities. Until something is known of the dental formula, or even of the structure of unworn cheek teeth, there will be no grounds for forming an opinion as to which of its known predecessors this form is most closely related. As Zangerl and Denison (in press) point out, *Astroconodon* extends the range of the Triconodonta from the Late Jurassic into the Early Cretaceous, a most welcome addition to our knowledge of the order.

One other, and very interesting, feature revealed by this latest triconodontine is that the group was by no means a static one, but underwent a considerable amount of change between the close of the Jurassic and the time of deposition of the upper part of the Trinity Sand. During this time, the evolution of the peculiar tongue and groove relationship of the molars must have gone on at a fairly rapid rate. This structure was certainly adaptive; every one of the diagnostic characters shown by the molars and last premolars is concerned in it. The reason for its development seems clear. The triconodontines, despite their great age, possessed, as Simpson (1933, p. 157) has remarked, one of the most perfectly carnivorous dentitions ever evolved. As he has further pointed out, the lower molars do not simply shear vertically past the uppers, but instead rake obliquely upward, inward and backward against them; great pressure must accordingly have been exerted on these

teeth and a correspondingly great strain put upon the lowers. Movements of individual lower molars would certainly have impaired the efficiency of this mechanism. Wood's review and discussion of individual tooth movement and interaction of adjacent teeth (1938) demonstrates that such phenomena are more widespread in mammals than is perhaps generally realized. The insertion and close fit of the tongue formed by the posterior cingulum cusp and the cement spur of each *Astroconodon* molar in the anterior groove of the tooth following would have prevented independent movement of the individual teeth of the series. The groove in the anterior root and the tongue formed by the ridge of the anterior alveolus wall would have almost entirely eliminated play of the series as a whole, which otherwise would have been possible due to the elasticity of the periodontal membrane (it may be noted in this connection that the roots of all cheek teeth of *Astroconodon* fall far short of filling their alveoli).[2]

It was mentioned above that the anterior groove of M_1 is as fully formed as that of M_2, whereas the posterior end of the last premolar is by no means perfectly fitted to it. Every biologist is continually impressed by the intricate interrelationship of parts of animals in structure and function. It is customary on the part of some, a practice perhaps more prevalent in the past than now, to proceed from a proper appreciation of such marvels of nature to the gratuitous assumption that natural selection is altogether too crude a mechanism to have brought them about. In *Astroconodon*, it seems to me that we have an example of such an intricate interrelationship in the making, but one not yet perfected by selection. There is evidence (Butler, 1939a; Patterson, 1949) that genetic factors affect the molar series as a whole; the development of the anterior groove in M_1 is in accord with this interpretation. This groove being present—ready and waiting, so to speak—any expression in the last premolar of the factors bringing

[2] The subject of individual tooth movement is one that has attracted little attention among paleontologists, the Wood brothers being notable exceptions. Aitchison has shown (1946) that the dagger-like canines of *Muntiacus, Hydropotes, Tragulus,* etc., which are used either for slashing from below or stabbing downward from above, have considerable freedom of movement within their alveoli, and that special adaptations are associated with such movement. This observation should be of particular interest to paleomammalogists, who are confronted with a considerable variety of creatures armed with such weapons.

about elongation of the posterior cingulum cusp would at least not be selected against. In *Astroconodon*, this cusp is definitely longer than in earlier genera but its complete adjustment to the groove has not yet come about. Later Cretaceous descendants of this form, if such there were, in all probability would have carried the trend to completion.

Mammalia of uncertain ordinal position
Gen. et sp. indet.

PM 543. Left humerus, incomplete. Data as for *Astroconodon denisoni* sp. nov.; in same pocket as PM 541.

The specimen consists of the greater part of the diaphysis, which, prior to burial, had suffered some breakage, chipping and abrasion. This is the fourth mammalian humerus thus far found in the Mesozoic and the first that has come to light in America. Despite its fragmentary condition, the bone yields a number of facts of interest.

MORPHOLOGY

The proximal end, as preserved, is not broken but is a somewhat abraded epiphysial surface that descends for a short distance on to the upper part of the delto-pectoral crest, precisely as in many later mammals. A typically mammalian epiphysis was obviously present. The shaft is somewhat bowed outward and is relatively robust compared to its length. The delto-pectoral crest is partially broken away but was evidently a prominent feature of the bone, extending antero-laterally well beyond the main body of the shaft. It terminates distally

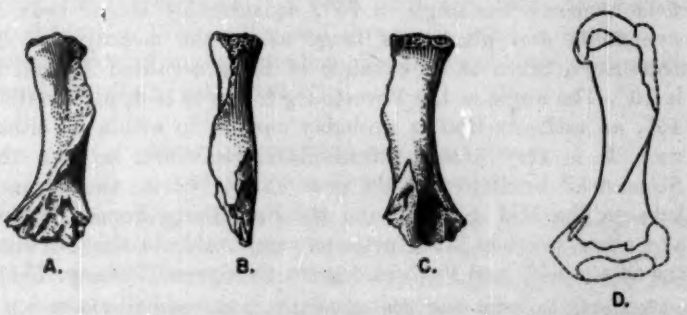


Fig. 3. Undetermined mammal of uncertain ordinal position. A, B, C, anterior, medial and posterior views of incomplete shaft of left humerus. D, tentative reconstruction in anterior view. X2.

at the center of the anterior face, two-thirds of the way down the preserved portion of the shaft. The parts preserved do not suggest an expanded delto-pectoral area but, rather, a single, fairly sharp crest. There is no clearly defined ridge distal to the lesser tuberosity, the rounded median border curving evenly over into the wide and deep bicipital groove. Vague markings and swellings in the appropriate position suggest that abrasion has removed attachment scars for the mm. latissimus dorsi and teses major. No trace of an entepicondylar foramen remains. Enough of the humerus is preserved to have shown some trace of this foramen if it had pierced a major portion of the distal end, as is the case in therapsids, monotremes and the Stonesfield humerus. If present, as is extremely probable, it must have been bridged by a relatively thin bar of bone, as in primitive Theria generally. The olecranon fossa extends far up the posterior face of the shaft, to a point almost on a level with the termination of the delto-pectoral crest; it is triangular in outline, so far as preserved, moderately deep and encloses a gently convex area medio-distally.

Although nothing remains of the epicondyles, enough of the distal end is present to show the general direction of the long axis of this part of the bone. The positions of the tuberosities are clear, and these reveal the general direction of the long axis of the proximal end. It is therefore possible to make a reasonable estimate of the angle between these axes and to compare it with the similar angle in the Stonesfield humerus on the one hand and in the therian mammals on the other. Simpson (1928a, pp. 154-160) has shown that in the Stonesfield humerus, the angle is 75°, considerably larger than in cynodonts and almost as large as in the monotremes; in *Didelphys*, taken as an example of the generalized Theria, it is 10°. The angle in the Forestburg humerus is approximately 45°, an estimate that is probably correct to within 5° either way. It is very nearly intermediate, therefore, between the Stonesfield[3] humerus and the generalized Theria; the advance between the Mid Jurassic and the Forestburg humeri in this and other respects being perhaps comparable to that between the Stonesfield[3] and Purbeck femora (Simpson 1928a, p. 154).

[3] It should be noted that this comparison, while morphologically valid, may have no phyletic significance. The Stonesfield limb bones may not be mammalian. Simpson (1938, p. 146) has noted that the humerus may have belonged to *Stereognathus*, now known to be the last recorded survivor of the Therapsida, and this possibility applies also to the femur.

The width of the bone across the distal end of the delto-pectoral crest is 3.84 mm.

DISCUSSION

As in the case of the Jurassic limb bones, the affinities of this humerus are quite uncertain. It was found, to be sure, in the same pocket as a triconodont jaw with no admixture of other mammalian remains, but to proceed from this to the assumption that the specimen is referable to the Triconodonta would be premature, to say the least. In the present state of knowledge, this can only be regarded as one of several possibilities. The heavy shaft of the Forestburg humerus, with its prominent delto-pectoral crest, presents a rather different appearance from the Jurassic humeri, which, in addition to the distinctions already mentioned, are more slender relative to length, and possess a well defined crest distal to the lesser tuberosity and a somewhat less prominent delto-pectoral crest. The multituberculatae humerus, known from the Late Cretaceous *Djadochtatherium* (Simpson 1928b, pp. 9-11) and the Mid Paleocene *Ptilodus* (Gidley 1909, p. 620; Simpson 1937, p. 92), differs in its more slender structure, less developed delto-pectoral crest, prominent crest below the lesser tuberosity, presence (in *Djadochtatherium*) of a third crest on the postero-median part of the bone, and less extensive olecranon fossa. There is nothing that would suggest reference of the Forestburg humerus to the Multituberculata or to the group or groups represented by the Jurassic humeri. Too much weight should not be attached to this purely negative evidence, however, for the humerus is a bone peculiarly liable to adaptive modification. Were moles and shrews, to take an extreme example, known from humeri alone, the degree of their relationship would hardly be suspected.

AGE AND CONDITIONS OF DEPOSITION OF THE MAMMAL-BEARING DEPOSIT

The very brief nature of the visits paid to the locality where the specimens were found allowed little time for stratigraphic observations. Fortunately, however, the geology of the area is fairly well known, so that the position of the specimens in the sequence can be established with reasonable accuracy.

The Trinity Sand (formerly known as Antlers Sand), the marginal transgressing facies of the Trinity Group, covers

most of the eastern part of Montague County. Bullard and Cuyler (1930) show it as having a gentle dip to the southeast, and describe it as "a fine, white to yellow pack sand occurring in massive beds 25 to 40 feet in thickness." In their published sections, they note cross-bedding and the occurrence of sands with many iron concretions. Well records reveal a thickness of some 600 feet (Bullard and Cuyler, 1930, p. 71). The sand at the Forestburg locality matches this account very well, as would be expected. Some cross-bedding was noted and there are innumerable iron concretions, ranging from pinhead size to irregular masses a foot or two in the largest dimension. Fossil wood is fairly abundant in the surrounding area.

In the easternmost part of the county, ridges formed by the Goodland Limestone and an equivalent of the Walnut Shale, both of the Fredericksburg Group, overlie the Trinity. Forestburg is situated on one of these ridges and the locality is a mile or so from its base, very nearly on the line of Bullard and Cuyler's section (1930, plate 3). A sight estimate, made at the time of the second visit, that the mammal locality is within 100 feet from the top of the Trinity Sand in this region seems reasonable.

As mentioned at the beginning of this paper, the fossil remains at the locality investigated are concentrated in small pockets, a yard or so across, all at essentially the same stratigraphic level. Apart from the presence of bone fragments, nothing was noted at the time that would distinguish the pockets from the surrounding sand. The fragments, many of them abraded, were presumably concentrated by current action. Mammalian remains are very definitely in the minority in these pockets, only three specimens having been recovered from among literally hundreds representing fishes, amphibians and reptiles. Some, at least, of the fishes — pycnodonts and small sharks—were marine, but the majority of the vertebrate remains are of land or freshwater types. The pockets clearly indicate near-shore deposition and suggest shore conditions similar in general to those at Stonesfield and Purbeck (Simpson 1928a), thus raising the hope that remains of other mammals, such as accompanied the triconodonts at these localities, will be found. An even closer comparison, perhaps, is with conditions in the German Rhaetic, as interpreted by Ehrat (1920, quoted by Simpson 1928a, p. 186; original

not seen), the only difference being that at the Forestburg locality, the vertebrate remains occur in pockets rather than in lenticular bone beds. The occurrence of numerous frog remains is a unique feature of the new locality. Their presence may reflect an unstable shore line and consequent rapid changes in the salinity of nearby freshwaters that were fatal for the frog populations living there. It may be noted in passing that some at least of these frogs appear to have been of leptodactylid affinities, a most important datum if confirmed by further study and, it is to be hoped, better specimens than those now in hand.

Stephenson *et al.* (1942) correlated the Trinity Sand with the early part of the Albian. The Forestburg locality is therefore considerably later than the Wealden, which is of Neocomian age. A glance at table 1 will reveal that this new

TABLE 1

Chart showing the relation of the Forestburg discovery to previously known Mesozoic horizons or localities that have yielded mammals (data from Simpson, Lull and Wright, Stephenson *et al.*, Parrington, Kühne). The German Rhaetic is omitted due to serious doubt concerning the right of the haramyids to inclusion in the Mammalia.

			North America	Old World
Cretaceous	Late	Maestrichtian	Lance Edmonton	
		Campanian	Belly River	Djadochta
		Santonian Coniacian Turonian Cenomanian		
	Early	Albian	FORESTBURG	
		Aptian Neocomian		Wealden
Jurassic	Late	Tithonian	Morrison	Purbeck, Tendaguru
		Kimmeridgian Oxfordian Callovian		
	Middle	Bathonian Bajocian		Stonesfield
	Early	Liassic		Fissure filling in Wales
Triassic		Rhaetic		Fissure fillings in England and Switzerland

discovery occupies a place in the section almost ideally intermediate in time between the comparatively well-known Purbeck and Morrison faunas and those of the Late Cretaceous. If therian mammals occur at Forestburg, significant advances in our knowledge of mammalian taxonomy and phylogeny should follow.

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Chicago, Illinois

[American Journal of Science, Vol. 249, January, 1951, Pp. 47-65]

ASSIMILATION IN THE COLD SPRINGS AREA OF THE WICHITA MOUNTAINS IGNEOUS COMPLEX, OKLAHOMA

JACK L. WALPER

ABSTRACT. In the Wichita Mountains igneous complex, near the village of Cold Springs, is an area which clearly illustrates the assimilation of one igneous rock by another. This assimilation of an andesite by a granite and the resulting hybrid rocks have been studied and described in detail. The hybrid products of the assimilation have a variable composition ranging from granite to quartz-diorite. Conclusions as to the assimilation are clearly shown by a study of field relationships together with petrographic study and chemical analysis.

INTRODUCTION

IN the vicinity of the village of Cold Springs, Kiowa County, Oklahoma (T. 4 N., R. 17 and 18 W.) there is an area of about six square miles of complex igneous geology consisting of intrusives and associated assimilation products. This forms a part of the igneous rocks of the Wichita Mountains.

Study of the Cold Springs area had as its objective the determination of the origin and emplacement of the igneous rocks and their age relationships. The problem was approached by three methods of investigation: (1) field mapping of boundaries of each igneous unit, study of field occurrences, contacts and relationships; (2) studies of thin sections; and (3) determination and comparison of heavy mineral content.

The field work was done during the school year of 1947-48 and in the summer of 1948. The field mapping was accomplished with the aid of aerial photographs from which overlay tracings of lineations and inferred outcrop patterns were made. All the quarries in the area were studied and specimens of the rocks were collected from selected localities.

Laboratory work consisted of gross examination of collected specimens and study of thin sections made from the same specimens. The heavy minerals of samples of the different rocks in the region were determined qualitatively and quantitatively for purposes of comparison and correlation.

The writer wishes to acknowledge his indebtedness to the Agriculture Adjustment Administration office in Hobart, Oklahoma, and to the Oklahoma Geological Survey in Norman, Oklahoma, for the loan of their aerial photographs covering

the area. Particular thanks is due T. S. Burkhalter for making the chemical analyses and to G. E. McKinley, who ably assisted in the field work. The writer is especially indebted to Dr. C. A. Merritt, under whose direction this paper was written, for the guidance and advice so freely given in connection with all phases of the work, and to Dr. W. H. Twenhofel for criticism in preparing the paper for publication.

PREVIOUS WORK

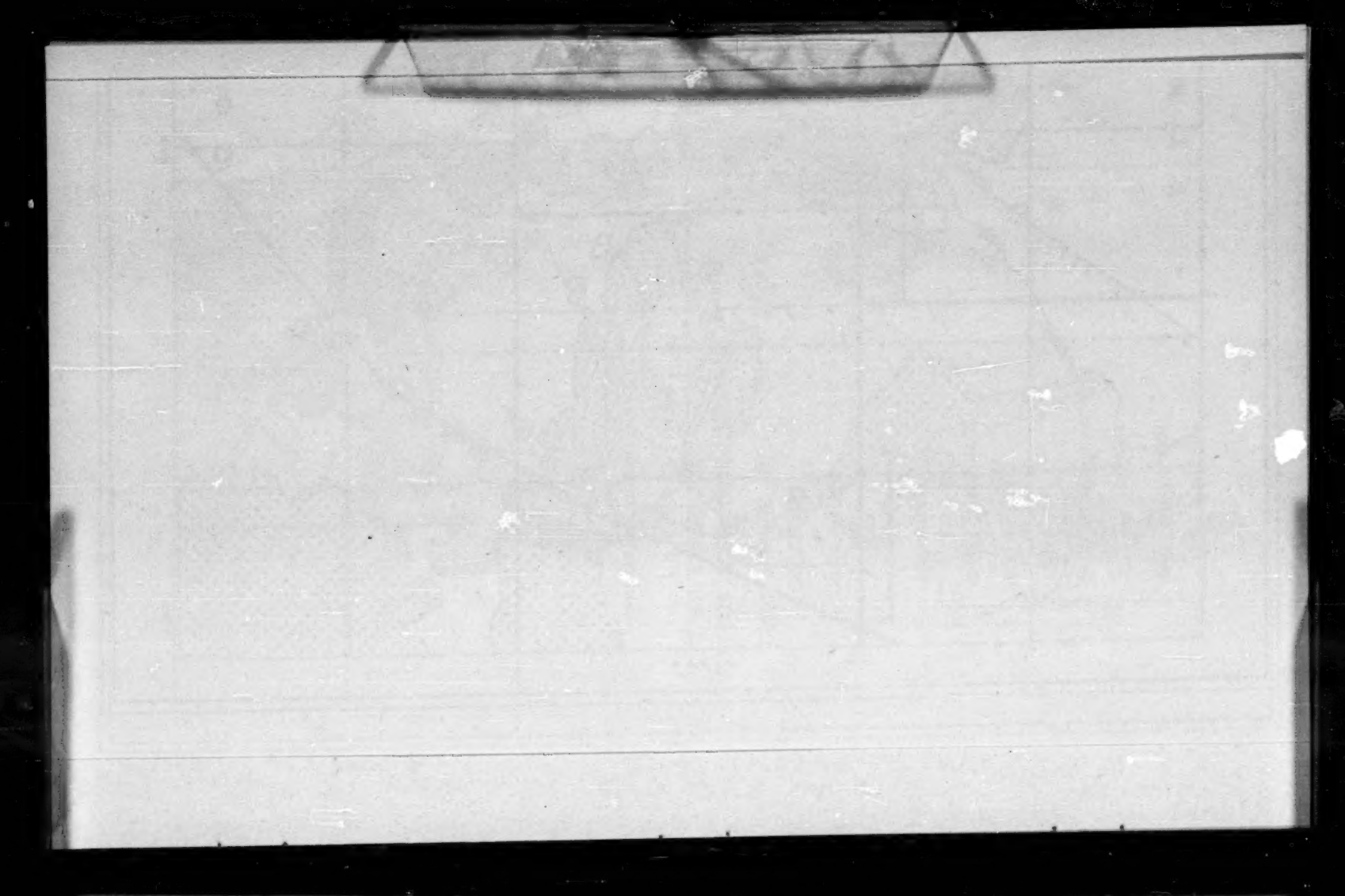
Early geological knowledge of the Wichita Mountains was obtained by various expeditions and geological surveys conducted at intervals during the latter part of the last century, among which were those of Shumard (1854), Comstock and Cummins (1889), Hill (1890, 1891), Vaughan (1899) and Bain (1900, 1904). Bain made a sketch map of the Wichita Mountain region and proposed the name Raggedy Mountains for the gabbro-anorthosite outcrops of the Roosevelt-Cold Springs area. Taff (1904) published a report and geological map of the Wichita and Arbuckle Mountain area.

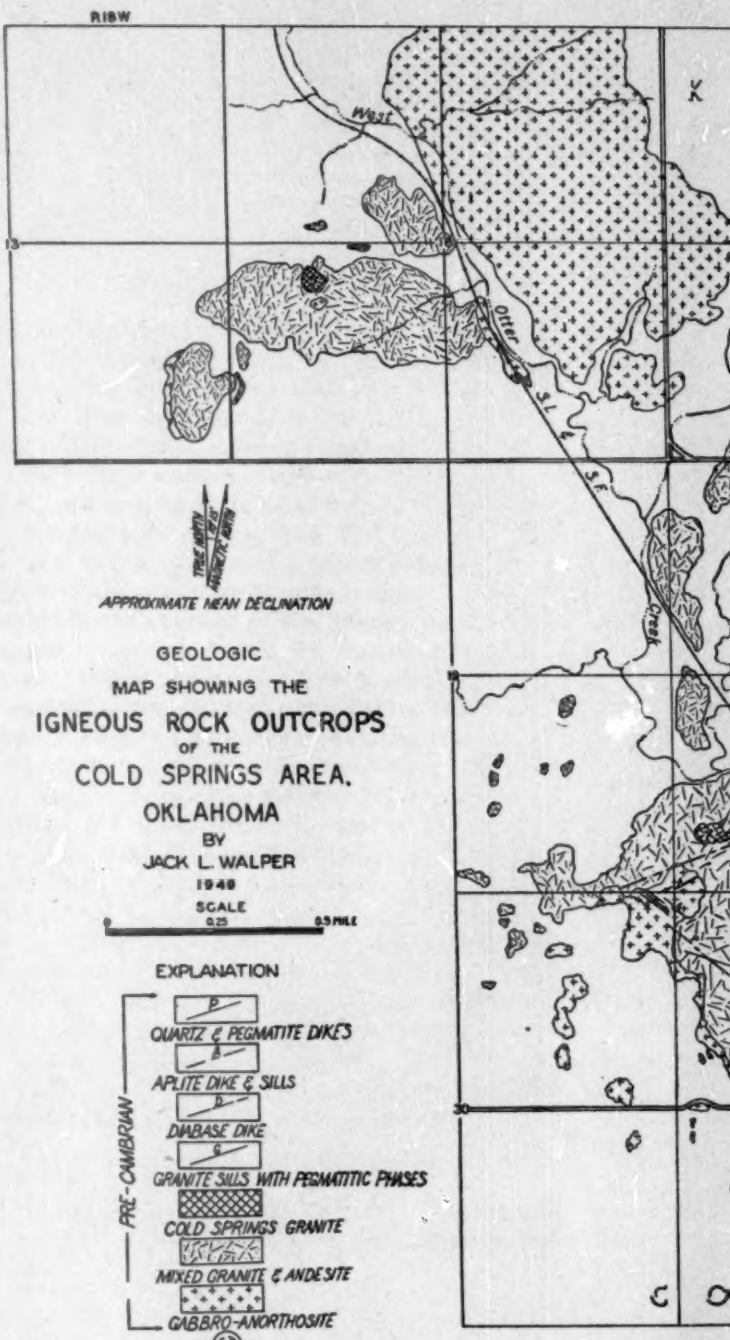
More recent studies of portions of the igneous rocks of the Wichita Mountains have been made by Taylor (1911 and 1915), Hoffman (1930), Anderson (1946), Scull (1947), Polk (1948) and Schoonover (1948); but Taylor and Anderson are the only geologists whose studies consider the igneous petrology of the Cold Springs area. Taylor divides the rocks of the Wichita Mountains into four groups: (1) pre-Cambrian sedimentary rocks, (2) pre-Cambrian igneous rocks, (3) Paleozoic sandstones and limestones of Cambrian and Ordovician age, (4) horizontally bedded, red sandstones and shales of probable Permian age.

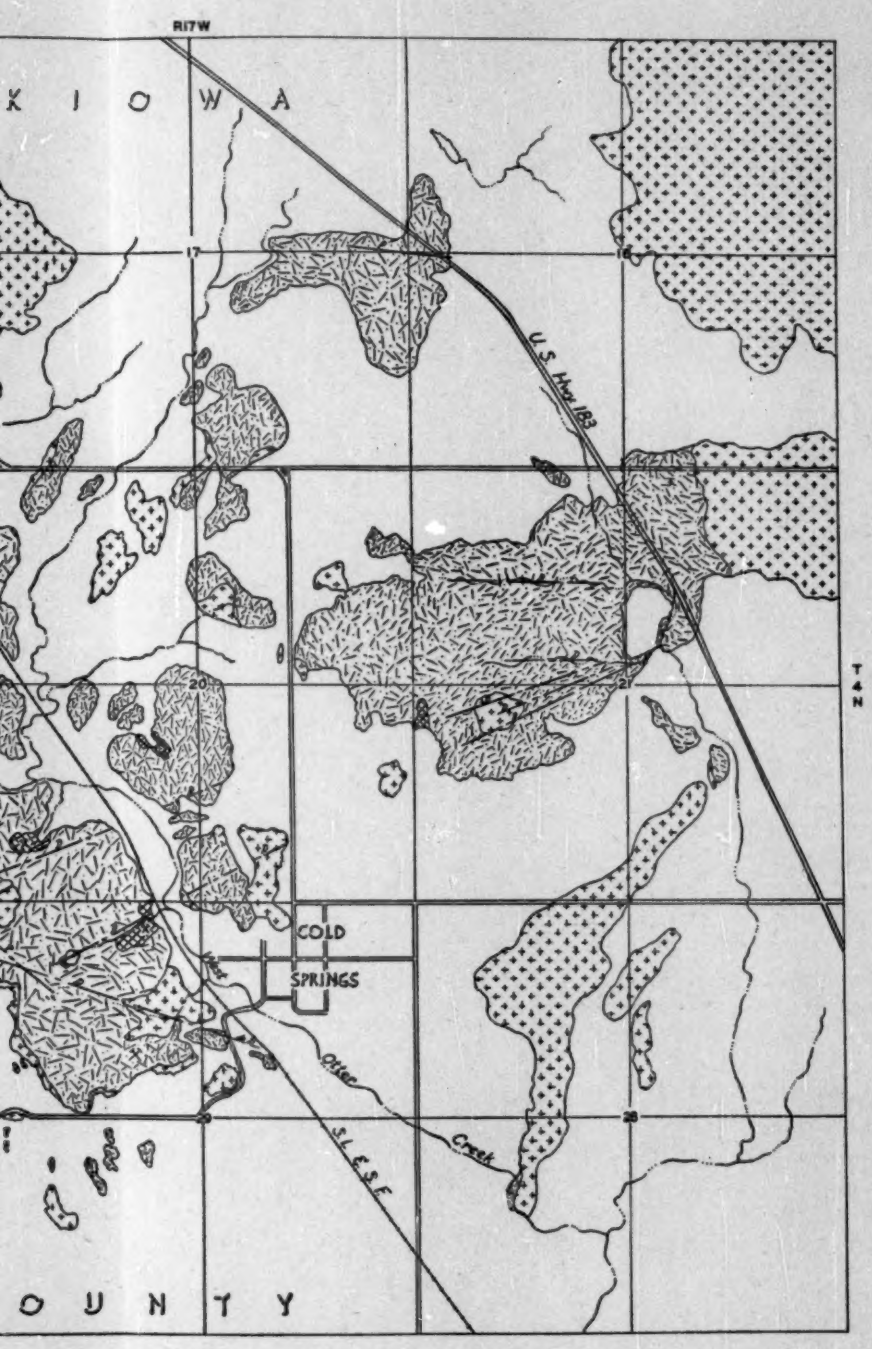
The pre-Cambrian igneous rocks of the Wichita Mountains consist chiefly of rhyolite, granite, quartz-monzonite, diorite and diabase. Taylor's (1915, p. 31) assigned age relationships of these rocks, reading from oldest to youngest, are: gabbro-anorthosite and Headquarters, Reformatory, Cold Springs, Lugert and Quanah granites. His second paper gives a detailed account of the dikes in the vicinity of Cold Springs.

TOPOGRAPHY AND VEGETATION

The type of rock controls the topography to a large extent. The resistant granites rise as rounded hills, moderately covered




R18W
Otter
S.F.
S.F.
Creek
APPROXIMATE MEAN DECLINATION
GEOLOGIC
MAP SHOWING THE
IGNEOUS ROCK OUTCROPS
OF THE
COLD SPRINGS AREA,
OKLAHOMA
BY
JACK L. WALPER
1948
SCALE
0.25
0.5 MILE
EXPLANATION
PRE-CAMBRIAN
QUARTZ & PEGMATITE DIKES
APLITE DIKE & SILLS
DIABASE DIKE
GRANITE SILLS WITH PEGMATITIC PHASES
COLD SPRINGS GRANITE
MIXED GRANITE & ANDESITE
GABBRO-ANORTHOSITE
QUARRY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O U N T Y

by bush or grass, above a grassy plain underlain by Permian sediments. The weathered dikes of the area often give rise to drainage courses.

Most of the area is barren of trees and those present are found along the water-courses and around small springs. The rather sparse character of the vegetative covering is largely due to the present semi-arid climate.

GENERAL GEOLOGY

Following is a resume of the major geologic events affecting the entire mountain system:

A series of igneous bodies, beginning with basic and ending with acidic composition, invaded ancient pre-Cambrian sediments. The only remaining vestige of these sediments is the Meers quartzite (Taylor, 1915, p. 32; Hoffman, 1930, p. 48; Merritt, 1948) which occurs as isolated roof pendants in the gabbro-anorthosite and Lugert granite.

The pre-Cambrian igneous activity throughout the Wichita Mountain system was followed by erosion. It is possible that early and middle Paleozoic strata, beginning with the Reagan (upper Cambrian) sandstone, were deposited in the Cold Springs area over the eroded surface, as they are known to have been deposited in some other parts of the Wichita Mountains. This period of deposition was terminated by a mountain-building uplift which formed the Wichita Mountains and is believed (Powers, 1928) to have had its beginning in early Pennsylvanian time and probably extended into early Permian. During this orogeny the region was uplifted, folded and faulted. Erosion stripped the sediments from the igneous core. Following this erosion, the sediments composing the Permian Red Beds were deposited over the upturned edges of the sedimentary rocks and over large areas of the then exposed pre-Cambrian igneous rocks. Later orogeny may have slightly affected the area.

Taylor (1915, p. 71) has discussed the vertical joints of the area and states that the prevailing direction for the major joints is east-west and north-south. There is also a set of horizontal joints which imparts a distinct sheet structure to the rocks (plate 1A). This jointing is very obvious in the Roosevelt quarry (Sec. 18. T.4N., R.18W.).

THE IGNEOUS ROCKS

The igneous rocks of the Cold Springs area are of pre-Cambrian age and include large bodies of gabbro, anorthosite, andesite, and granite, and dikes of aplite, pegmatite and diabase. The anorthosite, which is a phase of the gabbro and grades into it insensibly, is discussed under the heading gabbro-anorthosite. The andesite is younger than the gabbro-anorthosite and has intrusive relationships with it; still younger than the andesite and intruded into it is the granite. The intrusion of the granite into the andesite has given rise to a series of hybrid rocks of varying composition, among which are quartz-diorite and granodiorite. A series of aplite, pegmatite and dense diabase dikes penetrate the larger igneous bodies.

The map accompanying this paper shows the irregular nature of the exposures. The rocks were mapped as two main types: the gabbro-anorthosite and the granite-andesite mixture. This latter is mainly an intrusion breccia but in places the andesite has been assimilated and formed a fairly even-grained, gray granitic type, which Taylor named the Cold Springs granite. Where possible, an effort was made to map separately the areas where this granite hybrid is present. This proved very difficult, owing to the similarity of the weathered surfaces and the intricate relationship of the various rock types in the area. The quarry sites where this granite is of commercial quality proved to be the only places where distinction could be made with any degree of accuracy.

Hand specimens and the corresponding thin sections of all rock types exposed in the Cold Springs area were studied. These were deposited with Dr. C. A. Merritt at the University of Oklahoma, Norman, Oklahoma. In order to be consistent with recent studies completed in other areas of the Wichita Mountains, the Johannsen (1939, vol. I, pp. 141-161) classification was used in the microscopic description and naming of the igneous rocks.

Gabbro-Anorthosite

The gabbro-anorthosite is the oldest igneous rock known in the Cold Springs area and has the most extensive distribution. It is present on top of the Roosevelt Quarry hill in section 18, T. 4 N., R. 18 W., as very small isolated outcrops and

weathered boulders. The core of the hill forming the outcrop in sections 20 and 29, T. 4 N., R. 17 W. consists of this rock which outcrops at the top of the hill and again at its very base. The granite-andesite mixed rock forms a layer around the western side of the hill, resting upon the gabbro-anorthosite, The same conditions appear to exist on the hill in section 21, T. 4 N., R. 17 W.

The gabbro-anorthosite for the most part seems to weather much more readily than the other rocks of the area and the outcrops in places are strewn with cleavage particles of pyroxene crystals. This feature is well exhibited along the road into the quarry of the Oklahoma Granite and Monumental Company, just west of the village of Cold Springs. In several localities throughout the area the gabbro-anorthosite has a structure resembling schistosity. Taylor attributes this to local metamorphism but later investigators have ascribed it to magmatic flow.

A specimen of this rock, collected about one-eighth mile north from the SW corner of section 17, T. 4 N., R. 17 W. is considered to be fairly representative of the gabbro-anorthosite of the area. This rock varies a little in texture and grades from anorthosite to gabbro.

Megascopically, the fresh rock is medium- to coarse-grained, grayish-black in color and is composed almost entirely of dark-colored plagioclase feldspar showing bright cleavage surfaces. Most of the feldspar crystals are uniform in size and do not commonly exceed 4 mm. in diameter, although a few measure over an inch in diameter. These large crystals commonly resist weathering and in some cases form knobs on weathered surfaces. In many places knobs are not present but are replaced by spots of brown iron oxide.

Microscopically, the rock has an equigranular, holocrystalline, medium- to coarse-grained texture. From 85 to 90 per cent is composed of crystals of bytownite feldspar (Ab_{26} An_{74}), most of which exhibit albite twinning; some of the large crystals show schiller structure. The chief femic mineral is diallage which is in sufficient quantity to classify the rock as a gabbro. The accessories consist of magnetite, zoisite (non ferrian), olivine and calcite. In several places the feldspar seems to have been partly shattered and later cemented by the pyroxene, indicating the following order of crystalliza-

tion; magnetite (first), plagioclase, olivine, and then pyroxene. The zoisite and calcite are of secondary origin. On the basis of the Johannsen system of classification this rock is placed in class II, order 3, family 12, and named diallage gabbro.

Andesite

This rock is present as sill-like bodies in numerous places, and everywhere it is intruded by a network of dikes and stringers of granite, forming an "intrusion-breccia" consisting of xenoliths of andesite included in a granite matrix. Near the margins of the sill-like masses in which this breccia occurs, the andesite is relatively pure. In more central portions the andesite grades into granitic types of rock, which are believed to represent syntectic products produced by the intrusion and assimilation of the original andesite by the granitic magma. In places where the andesite is present in the granite as large blocks, or is cut by granite dikes, and these have been exposed to erosion, the contact between the two rocks has been eroded to form depressions on the surface. These vary in depth up to three inches. A fairly good example of this is shown in plate 1B.

Near the margins of the andesite bodies the rock is found to be relatively free from alteration by the granite. It was from this uncontaminated marginal facies exposed in the quarry in the NW part of section 21, T. 4 N., R. 17 W. that the specimen studied in this section was collected. Other slides of this rock showing contact relationships with the gabbro-anorthosite and granite were studied, as well as slides from the Taylor Collection at the University of Oklahoma.

Megascopically, the andesite is a dark-gray to black, aphanitic rock, holocrystalline and almost coarse grained enough to be classed as phaneritic. It is completely crystalline and not dense. The weathered surface has a brownish-green tint and a somewhat pitted character. The surface tint is due partly to the weathering of minute biotite flakes.

Miscroscopically, the rock does not show any glassy, felty or hyalo-ophitic texture common to aphanitic rocks of this family. Its minerals, although smaller in size, exhibit the same hypidiomorphic granular texture of the phanerites of this type. It has an ophitic[1] texture. The laths of plagioclase are

[1] Albert Johannsen, 1939, vol. I, p. 45. The word is used as Zirkel first used it in 1866. It applies only to the crystal arrangement and does not imply that the minerals must be plagioclase and pyroxene.

from 0.2 to 1 mm. in length and are andesine (Ab_{53} An_{47}). They are twinned according to the albite and carlsbad laws, which are commonly combined. The mafic minerals are for the most part hornblende and biotite. Minor quantities of pyroxene are present, some of which have altered to hornblende and chlorite. The hornblende is pleochroic in shades of green and brown, and contains magnetite inclusions. It has altered to biotite and chlorite in some places. The lath-shaped biotite is about the same size as the hornblende and ranges from .05 to 0.4 mm. in diameter. Both these femic minerals contain feldspar and magnetite inclusions, giving a poikolitic texture. Interstitial quartz containing apatite inclusions is present. Accessory minerals are magnetite and apatite. The order of crystallization is magnetite (first), pyroxene, feldspar, hornblende, biotite, apatite and quartz.

The rock is assigned number 2212E and named hornblende andesite.

Granite

The granite in the Cold Springs area has been mapped as Lugert, a granite present in other parts of the Wichita Mountains, by Taylor, but the correlation has not been established. It is pink in color and generally present as large and small dikes and stringers cutting the andesite. In places it has been injected into the marginal zone of the andesite in a series of small dikes and sills, while in others it has been intruded along the contact of the andesite and gabbro-anorthosite as large sills, which project past the contact and cut into the gabbro. In many areas these sills exhibit a pegmatitic texture, as may be seen in the sill remnants on the northwest slope of the hill in the northeast corner section 30, T. 4 N., R. 17 W.

Contacts between the rocks in the marginal zones are very sharp and any evidence of assimilation is lacking. Here the andesite is the most plentiful, with stringers of granite forming a matrix around large blocks of andesite. Conversely, as the marginal zones are left and one approaches the central regions of the sills, the junction between the two rocks grade from sharp to shadowy, due, apparently, to more complete assimilation by the granite. Here one can see many partly digested xenoliths of the andesite in a gray Cold Springs granite which is in contrast with the pink granite. In the more

central portions of the sill-masses of granite, the contacts with xenoliths of andesite become nearly obliterated and an occasional shadowy xenolith in the gray granite is all that can be seen of the original andesite.

Several samples of the marginal pink rock show it to be medium- to fine-grained, with a low mafic content. There are portions in the fillings between the angular fragments of andesite, which exhibit a pegmatitic phase.

Microscopic examinations were made of several slides. The pink rock has a hypidiomorphic granular texture and consists of orthoclase, quartz, plagioclase, hornblende, biotite, pyroxene and accessory magnetite, apatite, zircon and tourmaline. Chlorite, calcite, kaolinite, prehnite, leucoxene, hematite and limonite are present as secondary minerals. Orthoclase is the chief feldspar and occurs as subhedral crystals as well as in combination with plagioclase forming a perthite. A few lath-shaped crystals of plagioclase with albite-oligoclase composition are present; they are somewhat altered to kaolin, calcite and prehnite. Quartz occurs as cracked and shattered subhedral particles as well as fresh interstitial grains with apatite and tourmaline inclusions. The femic content varies a little and may be due to contamination of the granite by the andesite. Brown and green pleochroic hornblende is the principal femic mineral. Biotite and pyroxene are present and some alteration of these has produced chlorite. Accessories are mainly magnetite, apatite and zircon. The order of crystallization was probably magnetite (oldest), pyroxene, hornblende, biotite, plagioclase, accessories, orthoclase and quartz.

The rock number is 126 and the name is leuco-granite.

HYBRID ROCKS PRODUCED BY ASSIMILATION

Cold Springs Granite

A hybrid rock, known commercially as the Cold Springs granite, is found in several places throughout the Cold Springs region. An attempt was made to map the outcrop of this rock, but owing to its variable composition over small areas, its variable manner of weathering and its intimate relationship with the andesite and pink granite, this proved an exceedingly difficult task. Commercial quarries are the only places in which one can be certain of the type.

The main outcrops of the Cold Springs granite are the Roosevelt quarry, the Oklahoma Granite and Monumental

Company quarry, the Cold Springs Granite Company quarry, and a few small test quarries. The most important of these are: the lower of the two small quarries situated one-quarter of a mile south of the Oklahoma Granite and Monumental Company quarry and west of the village of Cold Springs; a boulder-blasting operation and small quarry located near center of section 20, T. 4 N., R. 17 W., on the west side of Otter Creek; the quarry farthest south of the two located on the east slope, near the south end of the large hill in section 21, T. 4 N., R. 17 W.; and the south-easterly one of the two quarries located just west of the highway in the NE corner, NW¼, section 21, T. 4 N., R. 17 W.

Taylor has adequately described this rock as a fine-grained gray hornblende-biotite granite of varying composition and texture. The light-gray surface upon close examination reveals specks of hornblende and biotite uniformly distributed through the fine-grained white and gray quartz and feldspar.

Microscopically, it contains varying quantities of orthoclase, albite, oligoclase, andesine, quartz, hornblende, biotite, magnetite, pyrite, apatite, sphene, and secondary chlorite, hematite, sericite and kaolin. Feldspar composes from 57.3 to 64.4 per cent of the rock. Both plagioclase and orthoclase are colorless or light-gray, and the plagioclase commonly penetrates or is completely enclosed by the orthoclase and quartz. The plagioclase ranges in composition from andesine (Ab_{53} An_{47}), which is badly altered to sericite, kaolin and prehnite, to oligoclase (Ab_{72} An_{28}). Taylor also reports albite. The oligoclase appears fresher than the more basic plagioclase and in some cases was identified as the plagioclase which forms the more acidic outer zone around the corroded more basic feldspar in the interior of the zoned crystals. Myrmekitic intergrowths are present along the edges of the plagioclase crystals but also form between the quartz and orthoclase. Brownish-green and green hornblende is present with green-to-black biotite. Accessories are mainly apatite, zircon, sphene, pyrite and sillimanite (?). The order of crystallization has been obscured by the injection of the granite magma. This interruption along with the deuteric or paulopost reactions in each of the magmas make the order of crystallization very difficult to determine.

The rock number is 226P and the name, due to the percentage of femic minerals and because Johannsen (1939, vol.

II, p. 124) prefers to omit the monzonitic types, is biotite-bearing hornblende granite.

Quartz-Diorite

Quartz-diorite is also a product of assimilation. Its main occurrences are (1) the railway right of way cut about one-half mile north of the village of Cold Springs, about 200 yards north of the Oklahoma Granite and Monumental Company's quarry, and (2) in the upper one of the two small quarries located south of the Oklahoma Granite and Monumental Company's quarry and west of the village of Cold Springs. In this last location it is associated with dikes and veins of granite. This rock represents the darkest and most basic assimilation product of the andesite by the granite.

This rock has a dark gray to black color with a mottling of lighter minerals. It has medium- to fine-grained equigranular texture and resembles the Cold Springs granite except that it is of much darker color.

In thin section the rock is hypidiomorphic granular and composed principally of feldspar, quartz, hornblende and biotite. The feldspar is a very basic andesine (Ab_{53} An_{47}) and occurs as laths exhibiting albite twinning lamellae. Many of the crystals have a zonal structure in which the outer zone is fresher and less corroded than the inner one. A part of the inner zone has a composition seemingly bordering on labradorite although this was not definitely established. The outer acidic zone in some of the specimens appeared to approach the composition of oligoclase when studied as crushed fragments in index oils. In some slides orthoclase feldspar is present as scattered crystals. Quartz occurs as subhedral crystals and rounded grains. Two ages of this mineral may be present. The older-appearing quartz occurs as fractured and shattered masses, whereas the younger, cleaner-looking material has been injected into the interstices between the feldspar crystals. In some cases there are feldspar fragments included in this supposedly later quartz. The femic minerals consist of hornblende, pyroxene and biotite. Hornblende was noted replacing pyroxene and also a dark brown biotite re-

Plate 1A. Horizontal or sheet jointing visible in a quarry near Cold Springs. Section 29, T. 4 N., R. 17 W.

Plate 1B. Veins and stringers of granite cutting andesite. This is common throughout the area, but especially well exhibited in NW part of section 29.

Plate 1A

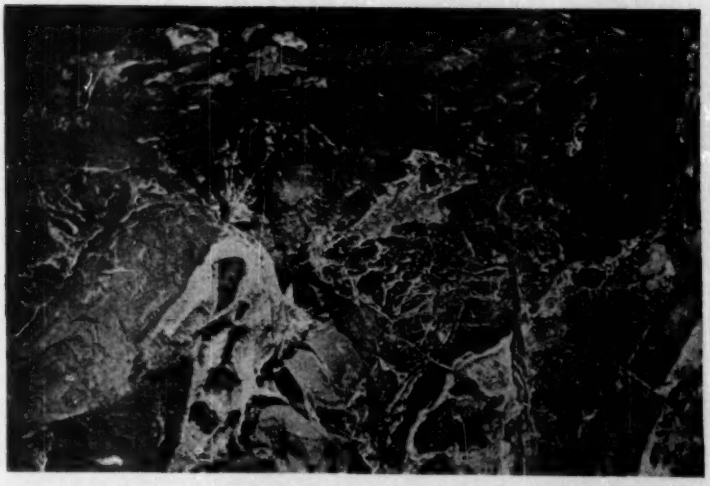

Plate 1B

Plate 2A

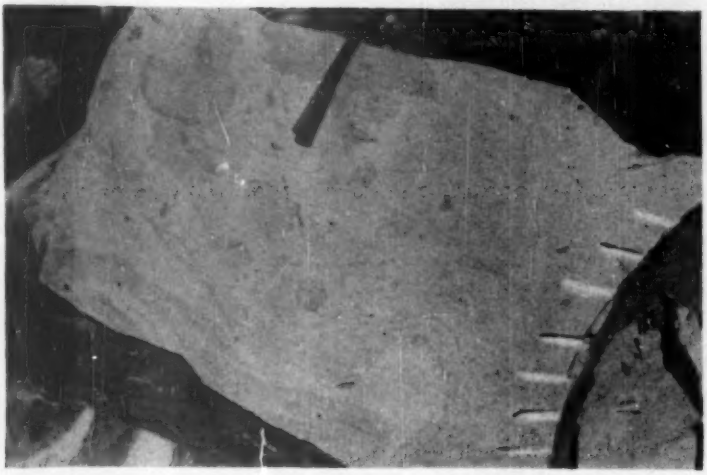

Plate 2B

placing hornblende. Early magnetite is commonly bordered or almost wholly replaced by hornblende and biotite. Some of the hornblende is twinned and the twinning lamellae are altering to chlorite; other hornblende is very much altered and eaten away, with chlorite the principal alteration product. Biotite ranges from fresh to slightly altered. The general appearance of the slides suggests paulopost reaction or exomorphic replacement.

The accessories are magnetite, apatite, zircon and sillimanite (?). The apatite is colorless and appears in a peculiar en echelon arrangement enclosed within the quartz. Some very fine needles of rutile may be present in some of the quartz crystals.

The rock is given number 228 and named tonalite or, more commonly, hornblende quartz-diorite.

Granodiorite

The granodiorite, also a product of the assimilation of the andesite by the granite, is intermediate in composition between the Cold Springs granite and the quartz-diorite, and grades imperceptibly into both. It is present in a small quarry on the east slope of the large hill in the west central part of section 21, T. 4 N., R. 17 W. In other quarries throughout the area where the Cold Springs granite is present, the granodiorite occurs as a phase of the Cold Springs granite.

Megascopically, this rock resembles the Cold Springs granite except for being of a slightly darker color. Microscopically, it is essentially the same as the Cold Springs granite, differing

Average Norms

Rock	Invading Gr.	Cold Springs Gr.	Grano-diorite	Quartz-diorite
Quartz	31.9	19.0	23.8	29.6
Orthoclase	49.7	36.2	25.4	3.2
Plagioclase	12.3	25.2	33.6	44.2
Hornblende	2.6	9.6	8.4	13.2
Biotite	2.1	7.3	7.5	7.1
Pyroxene	0.1	0.3	0.3	1.2
Magnetite	0.8	1.2	0.5	1.2
Apatite	Tr.	Tr.	Tr.	0.02
Zircon	Tr.	Tr.	Tr.	Tr.
Pyrite	—	Tr.	—	—

Plate 2A. Aplite dike in SW corner of section 20. It has resisted weathering and stands above the general surface.

Plate 2B. Flow lines and inclusions of andesite in Cold Springs granite. Small quarry west of Cold Springs.

only in the mineral percentages. It exhibits some of the same features as the quartz-diorite but is more acidic.

It is classified as number 227 and given the name biotite-bearing hornblende granodiorite.

DIKE ROCKS

Dikes of aplite, pegmatite, diabase, and quartz are common in the Cold Springs area. The aplite dikes are mostly very small and cannot be traced very far. The most prominent aplite dike is exposed on the top of the hill near the intersection of sections 19, 20, 29, and 30, T. 4 N., R. 17 W. It has resisted weathering better than the surrounding rocks and rises above the general surface (plate 2). This dike is vertical and is about three feet wide.

This flesh-colored, medium-grained rock has a xenomorphic granular texture in thin section. It is composed essentially of orthoclase and quartz with minor oligoclase and a few scattered crystals of hornblende. It is classed as a leuco-granite aplite and has the number 126.

The two best known pegmatitic dikes are in the central part of section 17, T. 4 N., R. 17 W., and on top of the hill in the central part of the NW¼ of section 29, T. 4 N., R. 17 W. Anderson (1946, p. 32) described the former, which penetrates the mixed granite-andesite. The latter is very similar and also cuts through the granite-andesite mixture, which makes up the greater portion of the hill where it occurs.

These dikes are composed essentially of quartz and orthoclase feldspar and crystals of the two minerals often are as large as 2 to 3 inches in diameter. The quartz is of the milky variety; locally it is graphically intergrown with the orthoclase.

Several diabase dikes have been found. These are usually highly weathered and produce a slight depression on the surface. This topographic expression is much more evident in aerial photographs than in the field and, consequently, it was from the photographs that many of the dikes were located and then examined in the field. Except in prospect pits, which are fairly numerous and usually located on dikes, the only field evidence of these dikes is the slight depression and particles of weathered rock fragments strewn about the surface along the "lead" of the dike. A typical example of this may

be seen on the road into the Oklahoma Granite and Monumental Company's quarry just west of the village of Cold Springs.

The map shows the main dikes. They range in thickness from two to five feet and have a vertical or nearly vertical dip. Since these diabase dikes weather very readily to soil that usually supports vegetation, it is probable that some were overlooked in the field.

The diabase dikes vary in color from a dark greenish-brown to a reddish-black, and are composed of very fine-grained, dense rock. Thin section examination revealed two types: one type, generally of a more brownish color, is highly weathered, and mineral identification was not possible; the other is microporphyritic and consists of phenocrysts of labradorite in a ground mass of what appears to be small feldspar laths and very badly altered femic minerals.

On the basis of what mineral identification was possible the rock was called diabase with number 2312.

The rocks have been cut by small quartz veins and stringers of variable thickness and shape. These appear to follow fractures and cut through the older rocks in many of the quarries, and also are present in the diabase dikes where they are associated with calcite veins. In some of the prospect pits where this was noted, the quartz and calcite veins are in a series forming a composite unit.

HEAVY MINERALS

The heavy minerals consist mainly of hornblende, biotite and magnetite, and nothing conclusive could be determined from a study of slides made from the separated fractions.

ASSIMILATION AS SHOWN IN THE ROCKS OF THE COLD SPRINGS AREA

Because of the discontinous surface outcrops of the igneous rocks in the area covered, as well as limited subsurface information, it is not possible to determine accurately the forms of the intrusions of all the rocks. The field occurrences and petrographic studies of the rock types show the Cold Springs granite, granodiorite, and quartz-diorite are not differentiates in their present form from one magma but represent syntectic products produced by the intrusion into, and cross-assimila-

tion of, the andesite by the granite. The term "cross-assimilation" is here used in the same sense set forth by Shand (1947, p. 67).

> It is clear that there are three processes by which foreign matter may be taken up by a liquid magma, namely:
>
> (1) Mechanical incorporation of solid particles without any chemical change;
>
> (2) Reactive solution and precipitation; involving the replacement of one solid phase by another solid phase;
>
> (3) Total dissolution, the solid phase disappearing altogether.
>
> It is very desirable that these three processes should receive distinctive and unequivocal names. Loosely speaking the term "incorporation" might be applied to all three of them, and "assimilation" to two; but much would be gained if we could agree to restrict "incorporation" to the first process, "assimilation" or perhaps "cross-assimilation" to the second and "dissolution" to the third.

Hence, cross-assimilation is the reactive solution and precipitation of a rock by a magma. This may result in profound changes in both the rock and the assimilating magma. There is a great deal of evidence that two of these geological processes, if not all three, have taken place in the igneous rocks of the present area.

The following is a review of the events taking place at the time of emplacement and also an attempt to establish evidence to substantiate the origin and emplacement of the igneous rocks in the Cold Springs area.

The gabbro-anorthosite is the oldest and most abundant rock and it is very probable that all outcrops of it are small parts of a large body which underlies the entire area. It is considered to be intrusive into pre-Cambrian sediments. There are no remnants of the original country rock in the Cold Springs area but it may be assumed that similar conditions took place here as in the Meers region several miles to the east, where some of the country rock remains. Merritt (1948) has stated that the gabbro magma intruded the original Meers sandstone and produced an unusual series of contact metamorphic rocks.

The size of the gabbro-anorthosite mass seems to be very large as Polk and McKinley found this rock extending far to the west. This extensive occurrence would support the idea that its outcrops in the report area are small parts of a batholith, sill, laccolith or lopolith; the exact form of the intrusion has not been proven to date.

Following the emplacement of the gabbro-anorthosite a more acidic magma was intruded into it, forming sill-like masses of andesite. The magma which formed this rock may represent a more acidic differentiate, at depth, of the gabbroic magma which formed the gabbro-anorthosite. From the remnants of the eroded surface outcrops, this intrusion seems to have formed, in most cases, thick lense-shaped or sill-like masses within the gabbro-anorthosite. The intrusion was accompanied by a contact border chilling which gave rise to a shell of solidified andesite around the sill-like masses.

At a relatively short interval of time after the intrusion of the andesite into the gabbro-anorthosite a granitic magma was injected into the andesite, but the intrusion was delayed long enough for the marginal zone of the andesite to become chilled. As a result of this chilling and early consolidation of the border facies, the peripheral shell of these sill-like masses was shattered and the granitic magma was injected into the resulting fractures. This magma engulfed the shattered fragments of the andesite, giving rise to an intrusion-breccia, with rounded to angular fragments of andesite cemented with granite. In the fractured marginal regions of the andesite bodies, the junction between the two rocks is very sharp indicating very little reaction. This chilled border phase, where the granite has formed a network of dikes and stringers in the andesite, would indicate that there was not sufficient heat in either of the rocks to cause cross-assimilation. Harker (1909, p. 357) has noted that the amount of mutual reaction depends upon high temperature in the solid invaded rock.

In the thick central part of the sill, owing to more prolonged flow and somewhat higher temperature, the acid magma has made greater inroads upon the earlier basic rock than it has in the thin part of the sill near the margin, and in places has totally destroyed it. This intrusion and cross-assimilation of the andesite by the granite has given rise to a series of hybrid rock types, varying in mineralogical composition and texture to such an extent that the differences are easily discernible in small exposures. The Cold Springs granite is an example of the rocks which make up this hybrid series, the quartz-diorite and granodiorite being other types intimately associated with it. Although only a minor part of the granite-andesite mixture, the Cold Springs granite is found as large

bodies in the central regions of what is left of the andesite sills, which remain only as small hills. The mixture, which has undergone only fracturing and injection without assimilation, forms a crust around the outer part of the hills and grades inward, becoming increasingly more granitic, into the Cold Springs granite.

The texture and mineral composition of the Cold Springs granite varies a great deal in different parts of the Cold Springs area. It is usually a fine-grained, gray, hornblende biotite granite of a wide color and composition range. In some places it is very light in color and closely resembles the intruded pink granite except for a greater mafic content. In other places, and in some cases in the same locality but in a different part of the quarry, it is darker in color due to a much higher per cent of hornblende and biotite. In these quarries unassimilated fragments of gabbro-anorthosite surrounded by flow lines were noted as well as shadowy xenoliths of the original andesite. This indicates incomplete assimilation of the andesite by the granite. These partly assimilated xenoliths and the andesite breccia are strong field evidence that this hornblende biotite granite, termed the Cold Springs granite, has been formed as a result of the intrusion and cross-assimilation of the andesite by the granite.

The pegmatitic phase of the granite associated with the basic xenoliths of andesite tend to support Bowen's (1922, p. 513) experimental conclusions that basic inclusions in a granitic magma or the addition of a basic magma to a granitic one advances it on its normal course of differentiation. This advancement in differentiation of the granite would explain the pegmatite and aplite phases which occur in the small dikes or apophyses between the inclusions of andesite. The pegmatitic granite is usually near the contact and there is a gradation to a finer-grained granite towards the center of the dikes. Emmons (1926, p. 423) has pointed out this same phenomenon in a study of inclusions in igneous magmas in several localities in Canada.

Microscopic examination of the Cold Springs granite has revealed further evidence as to its origin. The zoned feldspars are not proof but suggest that assimilation may have taken place. The presence of myrmekite intergrowths in this assimilated rock adds to the evidence of re-solution and pre-

cipitation of the andesite by the granite. Sederholm (1916) has described like intergrowths and shown that they are the result of the assimilation of a basic rock by a granite. Hogbom (1899) has observed the solution of a diabase in a granite at Ragunda, Sweden, and has pointed out not only the occurrence of a network of veins and dikes of granite cutting the diabase, a condition similar to one in the field of the present report area, but also the chemical and structural inconstancy of a series of intermediate rocks produced as a result of the assimilation.

Krieger (1932, p. 355) in a study of the zinc-lead deposits at Pecos, New Mexico, has also recognized that assimilation has taken place between a basic rock and granite. Besides field evidence he states as some of the petrographic evidence, the presence of myrmekite, the corroded centers of the zoned plagioclase feldspar crystals, and the replacement of hornblende by a deep brown biotite.

The thin section slides made from the syntectic rocks of the Cold Springs area exhibit many of these same features. This array of evidence strongly supports the conclusion that the Cold Springs granite and the other syntectic rocks, identified as quartz-diorite and grano-diorite, are the result of cross-assimilation.

In addition to the foregoing evidence, the chemical composition of the rocks concerned tends further to support the fact that they are cross-assimilation products. The hybrid rock should have a composition which can be calculated from the analysis of the two component rocks, taken in the proper ratio. The following table gives the SiO_2 and the CaO percentages of the rocks along with the calculated mixture:

Rock	SiO_2	CaO
Granite	74.0	0.9
Andesite	48.3	6.7
Calculated mixture	61.6	3.7
Cold Springs granite	61.6	3.2

A mixture of the granite and andesite in the proportions of 51.6 to 48.3 would give the SiO_2 percentage of the hybrid rock, or, in this case, the Cold Springs granite. It would also give 3.7 per cent of CaO. This calculated percentage is not much greater than that actually found in the hybrid rock. This slight difference may be accounted for by unequal dif-

fusion as has been pointed out by Harker (1909, p. 358) in a similar case of assimilation occuring in the British Tertiary plutonic rocks of Skye.

Therefore it seems probable that the Cold Springs granite, as well as the other minor syntectic rocks, namely granodiorite and quartz-diorite, are cross-assimilation products which have been produced by the intrusion, reactive solution and replacement of the andesite by the granitic magma.

To get a series of intermediate rocks as a result of this cross-assimilation seems to be the normal thing. Hatch and Wells (1937, p. 267) state that intermediate rocks of monzonite, tonalite and quartz-diorite composition are formed by interaction between acid magma and basic xenoliths. Krieger (1932, p. 361) also found some minor varieties of rocks in the Pecos area.

Following the emplacement of this series of rocks the region was later fractured and cut by a series of aplite, pegmatite and diabase dikes and quartz veins. There was considerable hydrothermal action associated with the diabase dikes. The igneous activity of the area was terminated by the formation of a few narrow quartz and calcite veins within the diabase dikes.

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[American Journal of Science, Vol. 249, January, 1951, Pp. 66-71]

BEDROCK SOURCE OF TILLS IN SOUTHWESTERN ONTARIO

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ABSTRACT. It has been found that the heavy mineral content of tills has been largely derived from crystalline rocks. The amount of light minerals of crystalline origin which must accompany the heavy minerals can be approximated. Hence, the amount of crystalline-rock material in the finer fractions of till can be established. The results indicate there is more such material in the sand and silt than in the gravel size.

INTRODUCTION

PLEISTOCENE geologists have long claimed that till is largely local in origin (Flint, 1947, p. 114; Salisbury, 1900). That is, the average distance of transport is only a few miles. This belief is based primarily upon the types of pebbles found in the till and little attention has been given to the finer fractions. Krumbein's work in Illinois shows the heavy mineral content of the till has been largely derived from a crystalline source. This is inferred from the abundance of unstable heavy minerals, such as horneblende, which could not possibly have come from the underlying sedimentary rocks.

The writer examined 25 heavy mineral suites from the tills of southwestern Ontario. The tills were in all cases underlain by non-crystalline rocks. The results paralleled those of Krumbein. It was found that the heavy mineral suites showed little variation and closely resembled those found in a granitic type rock. A similar observation is made by Kruger (1937) in a study of tills in Minnesota. It was, therefore, assumed that the heavy mineral content of the Ontario tills came from the Canadian shield.

The heavy minerals in the till must be accompanied, in some proportion, by light minerals of crystalline origin. Due to the small variation of the heavy mineral suites it may be inferred that the proportion of heavy to light minerals in different tills remains fairly constant.

Since the heavy minerals were probably derived from a mechanical breakdown of the coarser crystalline material in the till, it seems safe to assume that the finer crystalline material will be similar to that which is found in the pebble size. Therefore, by examining the pebbles it should be possible to approximate the proportion of heavy to light minerals in the finer

fractions. It has been mentioned above that the heavy mineral suites resembled those found in a granite. Therefore, if the amount of heavies is known and if a ratio of heavy to light minerals similar to that of a granite is used, it should be possible to determine the amount of crystalline material in a till.

PROPORTION OF HEAVY TO LIGHT MINERALS

The light mineral fraction of tills is made up of materials derived from both the Canadian shield and from underlying sediments. Examination of the light minerals showed that the predominant feldspars are orthoclase and microcline with minor amounts of plagioclase.

An average heavy mineral suite from southwestern Ontario is made up as follows: amphiboles 40-50 per cent, pyroxenes 10-15 per cent, opaques 8-15 per cent, garnet 12-20 per cent and the remainder made up principally of epidote, biotite, sphene, apatite, zircon, tourmaline, staurolite, kyanite, corundum, rutile and topaz. The types and proportions of heavy minerals coupled with the large amounts of orthoclase found in the light fraction suggest that the ratio of heavies to lights should be similar to that of a granitic type rock.

In a study of the Saganaga granites of the Lake Superior region (Tyler, Marsden, Grout and Thiel, 1940) it was found that the heavy minerals made up an average of 7-8 per cent by weight of the original rock. However, a laboratory examination of the crystalline pebbles in the tills of southwestern Ontario showed that 15-20 per cent were of a basic nature. The heavy minerals were tentatively assumed to comprise about 15 per cent of the total crystalline fraction of the till.

Samples near Lindsay, Ontario (fig. 1), disclosed a more accurate method of determining the ratio of heavies to lights and allowed a check on the value of 15 per cent used. In this area the drift rests upon Black River and Trenton limestones, which in turn lie directly on the Canadian shield. Since tills from this district contain only crystallines and limestones, insoluble residues should be a measure of the amount of crystallines present. As a precaution, the insoluble residues were examined under the microscope and it was found that over 90 per cent of the material was of a crystalline nature. The heavy minerals were then separated from the residues and it was found that they made up 13 per cent of the crystalline material, which is good

agreement with the above theoretical value. At present the percentage of heavy minerals in the crystalline fraction of the till is the main source of error, but it seems plausible that a value of 13-15 per cent is the right order of magnitude.

METHOD OF SEPARATION

Gravel Size (above 30 mesh screen size).—The pebbles were cracked open and identified according to rock type. The finer gravel sizes were identified under a binocular microscope and the crystallines separated.

Sand Size.—The sands were broken down into three size groups: (a) 30-100 mesh, (b) 100-170 mesh, (c) 170-250 mesh.

Fig. 1

The heavy minerals were separated using acetylene tetrabromethane (specific gravity 2.96) and centrifuging for an average of 25 minutes.

Silt Size.—It has been pointed out by Brown (1929) that any moisture left in the finer sizes has a great effect upon the separation of heavy minerals. It was found that driving off the moisture by normal evaporation or heating baked the silt, and hence the separation of the heavies was far from complete. The silt fraction was, therefore, dried in acetone and then brushed into the separating tube. By evacuating the separating tube while adding the heavy liquid, the yield of heavy minerals was increased by about 30 per cent over normal methods of separation.

Clay.—It was found experimentally that the percentage of heavies reached a maximum in the fine sand size and then tapered off into the silts and clays. Mechanical analysis of the tills in the Lindsay area, mentioned before, contained only small amounts of clay (5-8 per cent). However, the percentage of clay increased rapidly where the ice passed over the first shales. From these two lines of evidence it may be inferred that the clay is largely derived from the underlying sediments. Therefore, no attempt was made to calculate the amount of crystallines in this fraction.

Results.—The results are tabulated in table 1.

TABLE 1

Sample number	Sample location in southwestern Ontario	Percentage of crystalline-rock material in till	Approximate distance (in miles) to nearest crystalline-rock source	Percentage of crystalline rocks by pebble count
1	Lindsay	45.0	29	18-20
2	Oshawa	29.5	55	11
3	Whitby	26.0	59	14
4	Toronto	24.4	68	10
5	Hamilton	6.0	125	8
6	Simcoe	13.3	150	5
7	Port Stanley	13.0	174	4
8	Eagle	14.4	180	4

CONCLUSIONS

The results show that the till taken from Lindsay, Ontario, contains 45 per cent crystallines, yet a pebble count on the same till yielded only 18-20 per cent foreign material. In all cases, except one, there was more crystalline material in the sand and silt sizes than in the gravel size. It is, therefore, evident that the finer portion must be analyzed before the true nature of a till can be ascertained.

The apparent decrease in the proportion of crystalline material may be due to a fanning-out of the ice. Salisbury first stated this in 1900, but he attributed the local nature of the drift to lodgement and overriding by the glacier.

Finally, it is thought that the amount of material present in a till from a certain outcrop that the ice has passed over is a function of:

(a) fanning out of the ice

(b) area of outcrop exposed

(c) relative erodability of the outcrop

(d) possible losses due to plastering effects of sub-glacial till.

FUTURE WORK

(1) As pointed out earlier the main source of error to date is the ratio of heavy to light minerals. It is hoped that when more insoluble residues are analysed and when the percentage of heavies in the crystalline pebbles is more accurately ascertained a better proportion will result.

(2) Further work on vertical variations in till should give a better picture of glacial transport.

(3) This method may prove of some value in stratigraphic correlation.

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[American Journal of Science, Vol. 249, January, 1951, Pp. 72-77]

TREE TRUNKS PRESERVED IN A VOLCANIC FLOW IN THE NORTHERN CAMEROONS

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ABSTRACT. An unusual occurrence of casts of trees trunks preserved in a Tertiary flow of olivine basalt is described from the French Cameroons, the mode of preservation is discussed, and the few hitherto known phenomena of a similar nature are described. Finally, a record of the microscopic examination of the material is given.

A unique occurrence[1] of tree trunks in a late Tertiary basalt flow was recently discovered by the writer on the roadside between Meiganga and Ngaoundere, Northern Cameroons, 17 kilometers south of the latter. Embedded within a porous lava flow consisting mainly of black and brown-red (oxydized) scoria are several well-defined and some less recognizable petrified tree trunks with a maximum length of one meter, 20 cm. and a diameter up to 60 cm. What makes them so remarkable is the mode of preservation—though the tree shape is evident, the material consists of porous lava and no trace of organic matter or internal tree structure can be found (plate 1).

Within volcanic agglomerates and tuffs all the world over, the Cameroons included, remnants of plants and trees are not uncommon, the process of preservation more or less resembling normal sedimentation. In a true lava flow, however, such preservation is rare, although there are several instances of it.

Best known is Macculloch's upright tree (often called the "fossil tree of Burgh") at Kinlochscridain, on the west coast of Mull, Scotland. Nearly 5 feet broad and 40 feet high, the cast is largely filled with breccia, consisting of lumps of white trap and charred wood. The lower part encloses a partially silicified semi-cylinder of wood glittering with quartz crystals. A. C. Seward and R. E. Holttum describe this material as charred wood, very friable in texture, the tissues relatively

[1] The outstanding importance of this discovery was pointed out by the writer to the appropriate authorities at Ngaoundere, Yaounde, Brazzaville, and the Ministère de la France d'Outre Mer, Paris, with the request that in the interest of science the locality be fenced in. When the writer, accompanied by Monsieur Nickles, Géologue principal du Service des Mines, Brazzaville, visited the locality half a year later, the place had been turned into a quarry for road repairs by the Chef de Région of Adamaoua, and thus this rare phenomenon had ceased to exist.

soft and not petrified. Though the preservation is far from perfect, the annual rings being well marked but crushed, the tree definitely belongs to the comprehensive genus of *Cupressinoxylon* (coniferous wood).

If we consider the rapid destruction of organic remains by excessive heat from a body in a crematorium to any organic matter dropped into a blast furnace, it is obvious that the preservation of wood texture is due to exceptional conditions. The Geological Survey of Scotland report (1924, pp. 111-113) refers to a "prominent bed of red ash" close to the tree, while in the case of the Tavool tree, an analogue of Macculloch's

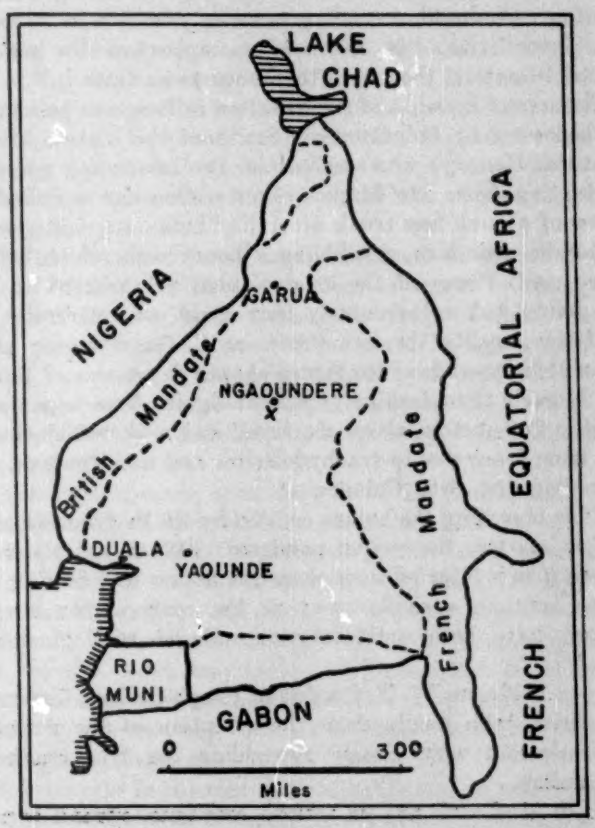

Fig. 1. Index map, showing location (X) of features discussed in text.

tree, near the north shore of Loch Scridain, Mull, the breccia is "rather more like an ash than a lava."

It seems to the writer that pre-existing explosive material, tuff or ash, acted as a "thermal cushion" neutralizing the heat effects in the same way as does a layer of asbestos around hot pipes. This theory is supported by the orientation of the basalt columns which, in accordance with the well-known laws of consolidation, arrange themselves at right angles to the surface of cooling. It is obvious that here a considerable "unit of low temperature" existed. The Survey report states in the case of Macculloch's tree "everywhere the columns are at right angles to the junction line and thus most arresting combinations are produced, including rosettes and fringes." And of the Tavool tree, "as the columns approach the mass of vegetable matter, they seem to tilt up away from it."

The second instance of preservation in lava was pointed out to the writer by Dr. Campbell Smith of the British Museum (Natural History) who showed him two interesting specimens originating from São Miguel, Acores. One was a cylindrical hollow of a small tree trunk which had been burnt away within porous basaltic lava, resembling a honeycomb, which had enclosed coal. Pieces of the original coal are believed to have been found, but unfortunately they could not be traced.

Of this locality the late Professor C. Gagel merely states that within the walls of the funnel-shaped depression of Furnas, São Miguel, thin, leaflike strata of lignite have been found, roughly 300 meters above sea level, and a threefold change-over from trachytes to trachydolerites and basaltic lava, thus resembling the Sete Cidades.

While observing the hollow created by the destruction of the organic matter, the writer pondered: "What exactly would happen if in a later effusive phase the hollow was filled in with molten lava, in a similar way as the casts of the dead of Pompeii have been artificially reproduced with plaster of Paris?"

It was Professor W. T. Gordon of King's College, University of London, who kindly drew the attention of the writer to a phenomenon very closely resembling the tree trunks of Ngaoundere.

Henry M. Cadell has on several occasions, spread over 35 years, described the occurrences of plant remains, including

tree trunks and branches, in an olivine basalt of Carboniferous age, "the matrix surrounding the fossil consisting of a crystalline massive volcanic rock," within the Bo'ness coalfield of West Lothian, Scotland.

According to the examining scientist, M. R. Kidston, the "outer surface might be described as an imperfectly preserved specimen of *Knorria*, which however might have originated from a *Lepidodendron* or a *Lepidophlois*."

With the exception of a thin coating of iron oxide and perhaps a little carbonaceous matter mixed with it, the upper part of the original plant is entirely replaced by crystalline calcite.

Later, after 1910, further specimens were obtained, one standing vertically in the whinstone. None of the specimens showed any internal structure: "They were externally sharply defined and composed internally of well-crystallized quartz and calcite, or of amorphous quartz in thin chalcedonic laminae, evidently deposited slowly in an empty pipe from which, after the lava had consolidated round it, all the original wood had entirely disappeared."

The same author found in the Museum of Morelia in the state of Michoacan, Mexico, a block of scoria or spongy basalt full of charred heads of maize, the shape of which was perfectly natural and unaltered. Some centuries ago a cultivated field of ripe maize in the neighborhood had been overwhelmed with lava which has cooled so quickly around the heads of corn that they had not had time to burn away before the matrix became solid.

Different instances, according to H. H. Cadell, have lately been recorded of lava streams on Etna, Kilauea and other volcanoes which have overflowed forests and consolidated around standing trees without destroying them. During recent eruptions of Vesuvius and Etna the lava has been seen to consolidate around erect trees in the same way. The correspondence with the tree trunk occurrences at Ngaoundere is nearly perfect. We must assume that the latter first formed a hollow caused by pre-existing wood having been burnt away; so far the origin is identical.

But whereas in the case of Bo'ness the hollow was infiltrated by mineral solutions, probably of a hydrothermal nature, during the post-volcanic phase, in the Cameroons the hollows were

filled by a later lava flow. When and in what manner the breaking-up process of the tree trunks took place, as they are found today in irregular chunks without orientation, is difficult to say. To the writer it seems likely that the whole process of being submerged under a lava flow necessarily involved a certain amount of dislocation.

Though silicified wood occurs abundantly within the Northern Cameroons, the expert examination of a thin section by Mr. P. M. Game of the British Museum (Natural History) rules out any possibility of such material having given rise to the tree trunks as found in their present state. Mr. Game determined a portion of the questionable tree trunks as a fine-grained, vesicular olivine basalt containing brown hornblende and showing a faint flow structure. A thin section showed the following minerals (in order of abundance):

Augite occurs in idiomorphic equidimensional crystals, giving characteristic eight-sided sections formed of the (100) (010) and (110) faces. The largest crystal is 0.4 mm. The mineral is faint brownish-grey and nonpleochroitic; several sections show twinning (sometimes repeated) on (100); optic axial angle is meduim to large and the sign positive; the mineral shows distinct dispersion.

Amphibole, a brown hornblende which forms idiomorphic to subidiomorphic crystals showing the characteristic prism faces and cleavages of the group, the acute angle of the prism cleavages being truncated by (010); cross sections are therefore six-sided. The crystals are commonly from 0.1 to 0.2 mm. in diameter (*i.e.* of the same dimensions as the pyroxene). The mineral is strongly pleochroic: a pale straw-yellow. $\beta = \gamma =$ deep reddish-brown. The optic axial angle is approximately 90 degrees. The extinction in vertical sections is straight or nearly so.

Plagioclase in laths averaging 0.1 mm. in length, showing slight parallelism in places, is difficult to determine exactly on account of its smallness, but is probably labradorite.

Olivine, though less abundant than the pyroxene and amphibole, forms relatively larger crystals — up to 0.6 mm. in length. These are mainly idiomorphic, with somewhat stumpy habit. Most of the crystals are perfectly fresh though sur-

Plate 1

Fig. 1. General view of lava flow containing tree trunks.

Fig. 2. Cross view of a section of tree trunk in lava flow.

rounded by a thin outer rim of iron oxide. The olivine is of a colorless magnesian variety with an optic angle of approximately 90 degrees.

These minerals are set in an opaque groundmass, extensively altered to hematite and pitted by roughly spherular vesicles from 0.1 to 0.2 mm. in diameter, which form from a quarter to a third of the whole rock. No infilling mineral was seen.

The writer gratefully acknowledges help and advice given by Dr. Campbell Smith and Mr. P. M. Game of the Mineral Department, British Museum (Natural History), London; Professor W. T. Gordon of King's College, University of London, and Professor H. H. Read of the Imperial College of Science and Technology, London; also Dr. N. R. Junner, Consulting Geologist, Selection Trust.

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REVIEWS

Bibliography of Organic Reefs, Bioherms, and Biostromes; edited by W. E. Pugh. Pp. xxxii, 140. Tulsa, 1950 (Seismograph Service Corporation).—This bibliography was "compiled as a service to the oil industry by the Seismograph Service Corporation," and will also be useful to many other geologists. It lists well over 1,000 articles in an author bibliography, supplemented by and cross-referenced to a geographic and a stratigraphic index as well as a short subject index. It is restricted to articles written in English and in large part to those published in North America. No attempt has been made to annotate the bibliography or indicate the relative importance of the references; major contributions are listed cheek by jowl with three-paragraph announcements of new oil fields in the trade journals. Annotation or selection might, however, have been invidious and would have vastly increased the already formidable job of preparing the bibliography. The staff of the Seismograph Service Corporation and of the department of Geology at the University of Tulsa, and especially Mr. Pugh, are to be heartily thanked for successfully carrying out this project. JOHN RODGERS

Introduction to Theoretical Igneous Petrology; by Ernest E. Wahlstrom. Pp. vii, 365. New York, 1950 (John Wiley & Sons, Inc., $6.00).—This book fills a need in the existing literature of igneous petrology. It is divided into two parts. Part I consists of four chapters, treating mainly of heterogeneous equilibrium, the phase rule, and equilibrium in silicate systems. Part II, consisting of ten chapters, takes up the major problems of igneous rocks. Several chapters are devoted to magmas, their source and nature, their emplacement in the crust, their crystallization, differentiation, and assimilation. Chapter 11 takes up the controversial theme of granitization, and the remaining chapters deal with epimagmatic processes, origin of rock types, and classification of igneous rocks, Each chapter is buttressed with a list of select yet comprehensive, up-to-date references. Finally, there is an appendix of 14 pages—Physical-Chemical Concepts Useful to Petrologists—in which the fundamental concepts of physical chemistry necessary for a quantitative understanding of petrology are briefly presented. The treatment is more condensed than it is in standard texts of physical chemistry, which is probably an undeserved compliment to geological students. Examples drawn from petrology would be helpful in enforcing the value of the principles. The writer is to be congratulated on his organization and synthesis of material in this relatively new mode of presentation. ADOLPH KNOPF

Volcanological Observations; by FRANK ALVORD PERRET. Pp. xi, 163; 117 figs. Carnegie Institution of Washington Pub. 549. Washington, D. C. 1950 (Paper cover, $5.00, cloth bound, $5.50). —The present work, as the author himself says, is a condensed record of observations made during some forty years of more or less continuous work on volcanoes the world over. The author studied at first hand the great outburst of Vesuvius in 1906 and successively, eruptions at Stromboli and Etna in Italy, at Teneriffe in the Canary Islands, at Kilauea in the Hawaiian Islands, and at Sakurashima in Japan. He studied the eruption of Mont Pelee in 1930, and the volcano-seismic crisis at Montserrat from 1934 to 1937. The book, which is divided into eight chapters, begins by describing the construction of the volcanic edifices and ends in a short chapter on diagnosis and prediction of volcanic outbreaks and their probable duration. The treatment deals exclusively with the observable facts and largely eschews speculative matter. The book can profitably be used as an introduction to volcanology. As it is essentially the record of the author's investigations, little attention is accorded to the work of other investigators.

An arresting feature of the book is the superb photographic documentation which vividly brings to the reader unsurpassed visual conceptions of the wide range of volcanic phenomena. Three notable monographs were the outcome of the author's earlier work, and this monograph, the fourth and last, as Leason H. Adams writes in the foreword, "will stand as an enduring monument to a great volcanologist." ADOLPH KNOPF

The Supersensitivity of Denervated Structures; by W. B. CANNON and A. ROSENBLUETH. Pp. x, 245; 62 figs. New York, 1949 (The Macmillan Company, $5.50).—When Walter Cannon died in the fall of 1945 he left among his effects the partly completed manuscript of a book on what he had termed some years before a "law of denervation." This half-written work has now been completed by Dr. Rosenblueth. As the senior author's long-time collaborator and erstwhile colleague during the last decade of a brilliant period for physiology at the Harvard Medical School, he is eminently fitted to do so. It seems somewhat anomalous in view of this that Dr. Rosenblueth's current collaborator, Norbert Wiener, speaks so little of Professor Cannon's contribution to self-regulating physiological mechanisms. Homeostasis, as he called such steady state regulation, was an important sequel to Claude Bernard's much earlier emphasis on the constant internal environment of animals. Together these concepts have been most fertile for physiology and biochemistry. Historically this relation between Bernard and Cannon is also paralleled by the

former's intuitive generalization of an earlier law of denervation, foreshadowing the one here presented in detail.

Approximately one quarter of the book has been written directly by Dr. Rosenbleuth, one half by Dr. Cannon, the rest jointly. The result, however, is not a patchwork but a smoothly organized whole.

Divided into five sections, the book begins with a description of the earliest pertinent experiments and then proceeds to define its scope. In the second part there ensues an enumeration of the data on muscle, melanophores, glands, and neurons as they relate to supersensitivity to chemical agents after denervation. Similar increased sensitivity to nerve impulses and some aberrant but perhaps similar phenomena are discussed in the two succeeding sections. The final chapters consider the nature of various hypotheses which might account for the facts and analyze their relation to certain experimental and clinical problems. Over 250 references are cited, including 54 since 1940 but only a few after 1945.

Throughout, the presentation is clear and well illustrated. Theoretical considerations are presented strictly as such and perhaps are somewhat disappointing in their tentative nature. But this constraint is forced by the complexity of the phenomena involved and the patent need for further research before more definite mechanisms may be profitably hypothesized. The data of this monograph should prove stimulating to thoughtful workers in many fields: the embryologist interested in the functional differentiation of nerve and muscle, the physiologist concerned with basic phenomena in excitable tissue, and the clinician for many practical problems. The comparative physiologist will regret that the material is mainly limited to vertebrate and even mammalian systems but should find many provocative ideas to follow up and perhaps illuminate in other animals.

Scientists will be relieved to learn that, as the foregoing suggests, the editorial irresponsibility recently demonstrated by the Macmillan Company in issuing one of its current best sellers as a contribution to science is not at all evident in the monograph under review. It is equally certain, and perhaps no more surprising, that "The Supersensitivity of Denervated Structures" will arouse but modest enthusiasm in its publisher's promotion and business offices. Nor will it provide much fodder for popular magazines or Sunday supplements. Nevertheless, it is an excellent book, a welcome addition to Experimental Biology Monographs, in which series the same authors' "Autonomic Neuro-Effector Systems" is a well-known classic.

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PAYNE & LANE, *Printers* — NEW HAVEN, CONN.